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Abstract

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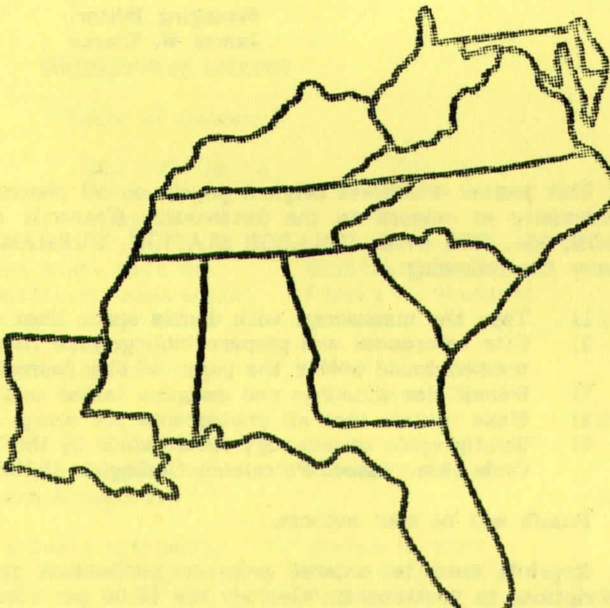

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SOUTHEASTERN GEOLOGY

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TRANSITION FROM EASTERN SLATE BELT TO RALEIGH BELT IN THE HOLLISTER QUADRANGLE, NORTH CAROLINA

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ABSTRACT

The Hollister 7 1/2-minute quadrangle in the northeastern Piedmont of North Carolina lies astride the boundary between the Eastern Slate belt and the Raleigh belt. Rocks of the southern portion of the area include middle to upper greenschist facies metagraywacke, metamudstone (argillite and phyllite), metasiltstone, metaconglomerate, mafic to intermediate metamorphosed intrusive rock, metabasalt and pre-metamorphically altered intermediate and felsic metavolcanic rock. The northwestern third of the quadrangle is dominated by pelitic schist, micaceous quartzite, amphibolite and greenstone of the lower to middle amphibolite facies. In the extreme northwest corner of the quadrangle, mylonitic microcline-rich orthogneiss is juxtaposed against the potassium feldspar-free metapelites. The boundary between the greenschist facies rocks and amphibolite facies rocks is a steep, apparently continuous metamorphic gradient. The distance along the present erosion surface from the biotite-in isograd to the staurolite-in isograd of pelitic rocks is approximately eight kilometers. Where outcrop permits, increasing metamorphic grade is continuously traceable in most of the lithologies. All the rocks had a common suite of protoliths and all may be included within a volcanogenic depositional sequence characteristic of the Eastern Slate belt. The only lithologic unit in the area of continental affinity (basement?) is the amphibolite facies felsic orthogneiss in the extreme northwest. In the Hollister area at least, the lithologic boundary between the volcanogenic cover sequence, of possible ensimatic origin, and ensialic basement does not coincide with the metamorphic transition from relatively low to relatively high-grade rocks.

INTRODUCTION AND GEOLOGIC SETTING

This study was undertaken in order to determine the nature of the boundary between greenschist facies metamorphic rocks of the Eastern Slate belt and amphibolite grade rocks of the Raleigh belt. The Raleigh belt-Eastern Slate belt boundary is an important feature which must be considered in any tectonic interpretation of the pre-Mesozoic history of the eastern Piedmont. Determination of the nature of this boundary is therefore of value in placing constraints on tectonic models dealing with this part of the southern Appalachians.

The belts of the southern Appalachian Piedmont (Figure 1) historically have been differentiated according to relative metamorphic grade (King, 1955 and numerous other authors). In North Carolina, both the Carolina and Eastern Slate belts contain volcanic and sedimentary rocks metamorphosed to greenschist grade while the Raleigh belt comprises gneisses and schists of the amphibolite facies. Although the Raleigh belt-Eastern Slate belt boundary has received little attention, the contact between the Raleigh belt and Carolina Slate belt to the west has been variously interpreted as an unconformity (Parker, 1977), a metamorphic gradient (Parker, 1968; 1978; 1979) and a fault boundary (Farrar, 1985a; Wylie, 1984; Wylie and Stoddard, 1984). It should be noted that these apparent discrepancies reflect differences in belt definitions as well as differences in interpretation of the geology. Defining the belts on the basis of metamorphic grade may be of little use in con-

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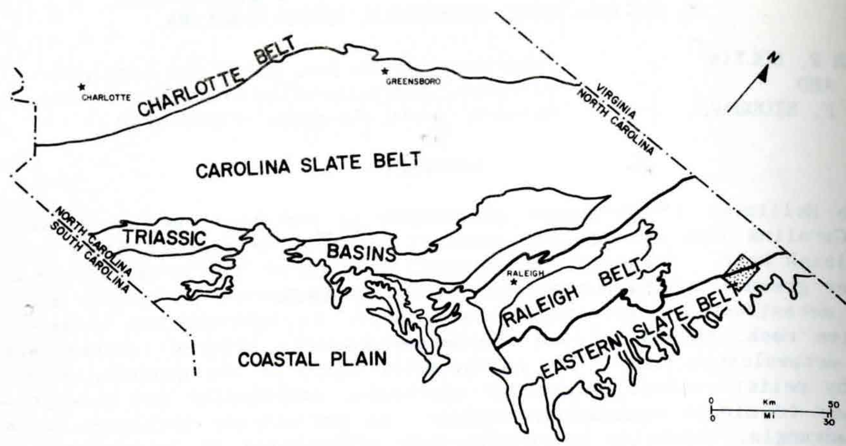

Figure 1. Geologic index map of the central and eastern North Carolina Piedmont (adapted from Williams, 1978). Location of the Hollister 7 1/2' quadrangle (Figure 2) is indicated by the stippled box. Belt boundaries are approximate only.

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structing tectonic models, because in many cases the regional metamorphism is a late event superimposed on more fundamental lithotectonic terrane boundaries (see Farrar, 1985a; Russell and others, 1985). In an attempt to avoid confusion, in this paper we will refer to the *metamorphic* Raleigh and Eastern Slate belts when using the terms in the old sense, and to the Raleigh and Eastern Slate belt *terranes* when discussing premetamorphic lithotectonic characteristics of the various units. When no qualifier is added, either of the two definitions is appropriate.

The metamorphic Raleigh belt coincides approximately with the Wake-Warren anticlinorium, which contains the late Paleozoic Rolesville granitic batholith in its core. The batholith has low and flat regional magnetic and gravity signatures, in contrast to the more variable patterns of the stratified rocks surrounding it to the east, west and south (Farrar, 1985a). Previous work in the eastern Piedmont includes that of Parker (1968), in which he produced a preliminary structure map. He defined a possible domal structure in the Hollister area. More recent reconnaissance mapping projects have been carried out in the area by the North Carolina Geological Survey (McDaniel, 1980; Wilson, 1981).

The most recent and detailed reconnaissance mapping in the eastern Piedmont is that of Farrar (1980; 1985a, b). His work defined a series of regional-scale folds and major ductile shear zones which divide the eastern Piedmont into tectonic blocks. Most of the Hollister area occurs mainly within his Eastern Slate belt block, which is separated from the Raleigh block on the west by the Macon mylonite zone and from the Roanoke Rapids block to the east by the Hollister mylonite zone (Farrar, 1985a). According to Farrar, these mylonite zones formed on the attenuated limbs of the Spring Hope synform during the Alleghanian orogeny. He also reports the occurrence of an earlier (pre-Alleghanian?) regional decollement which is the contact between a volcanogenic cover sequence represented by rocks of the Carolina and Eastern Slate belts, and proposed Grenville-age (Farrar, 1984) continental basement of the Raleigh block. The decollement is interpreted by Farrar to have been overturned in the Macon mylonite zone.

The Hollister mylonite zone is an important constituent of the Eastern Piedmont fault system of Hatcher and others (1977) who, on the basis of geophysical data, suggest it is an extension of the Augusta fault of South Carolina and Georgia. The latter juxtaposes the Belair and Kiokee belts, possible respective equivalents of the Eastern Slate and Raleigh belts.

Finally, a number of mafic and ultramafic associations have been observed within Slate belt lithologic units both east and west of the Raleigh belt terrane. The Halifax County complex is a sequence of such rocks which outcrop in southwest Halifax County, just southeast of Hollister. The complex is inferred to be a partial ophiolite sequence (Kite and Stoddard, 1984; Stoddard and others, 1982).

STRATIGRAPHY OF LAYERED SEQUENCE

Metavolcanic and metasedimentary rocks in the Hollister area (Figure 2) belong to the Spring Hope formation, defined informally by Farrar (1985b) to include the following three members: phyllite-metasiltstone, felsic metavolcanic rocks and mafic metavolcanic rocks; the formation can be further subdivided in the Hollister area. These strata constitute a metamorphosed sequence of interlayered graywacke, conglomerate, siltstone, mudstone, mafic to felsic volcanic rocks and possible impure limestone or marl. No fossils have been found in these rocks, but discoveries in the Carolina Slate belt (St. Jean, 1973; Secor and others, 1983; Gibson and others, 1984) of North and South Carolina and the Belair belt (Maher and others, 1981) of South Carolina and Georgia suggest that, if the rocks in the Hollister area are broadly correlative, the probable age of deposition was between late Precambrian and Ordovician time.

This stratigraphic section is in contact with felsic orthogneiss along the Macon mylonite zone in the northwest corner of the Hollister quadrangle. Mylonitic fabric is strong in the gneiss. Ages for the high-grade orthogneisses and related rocks in the central Raleigh belt have not been determined but it is suggested that they may be Grenville (Farrar, 1984). The orthogneiss belongs to the Macon formation of Farrar (1985b); it will be described first, then the remainder of the stratigraphy will be described in order from the inferred oldest units to youngest.

Felsic Orthogneiss Unit

Outcrops of felsic orthogneiss are found in the northeast corner of the Inez quadrangle, located immediately west of the Hollister quadrangle (Figure 2, sample location 1; see also Appendix). The north-northeast strike of the foliation parallels that of the rocks to the east. The gneiss is granitic in composition and contains quartz (20-40%), perthitic microcline and orthoclase (20-30%), oligoclase (10-15%), and minor amounts (less than 10% each) of muscovite, biotite, epidote and opaque minerals. Potassium feldspar and plagioclase form relatively resistant porphyroclasts which sit in a matrix of fine-grained, recrystallized quartz, feldspar, platy oriented mica, and sparse equant grains of epidote and opaque minerals. Quartz also occurs in sheared-out ribbons.

In outcrops of the felsic orthogneiss examined, feldspar prophyroclasts display asymmetric tails of finer-grained, recrystallized material that consistently show a dextral sense of shear (Figure 3) as determined using the criteria of Simpson and Schmid (1983; see also Simpson (1986)). Thin bands of pelitic composition are interlayered locally with the orthogneiss, and massive hydrothermal quartz veins occur along the contact with the potassium feldspar-poor pelitic schist unit to the east.

Pelitic Schist Unit

Quartz-muscovite schist and quartz-poor muscovite schist comprise a pelitic unit which grades stratigraphically upward (eastward) into tuffaceous metasiltstones and metagraywackes. The metapelites interfinger with amphibolites and are juxtaposed against the mylonitic felsic orthogneisses along the Macon zone to the west. Good exposures of schist are found near the western boundary of the Hollister quadrangle (Figure 2, location 2; see

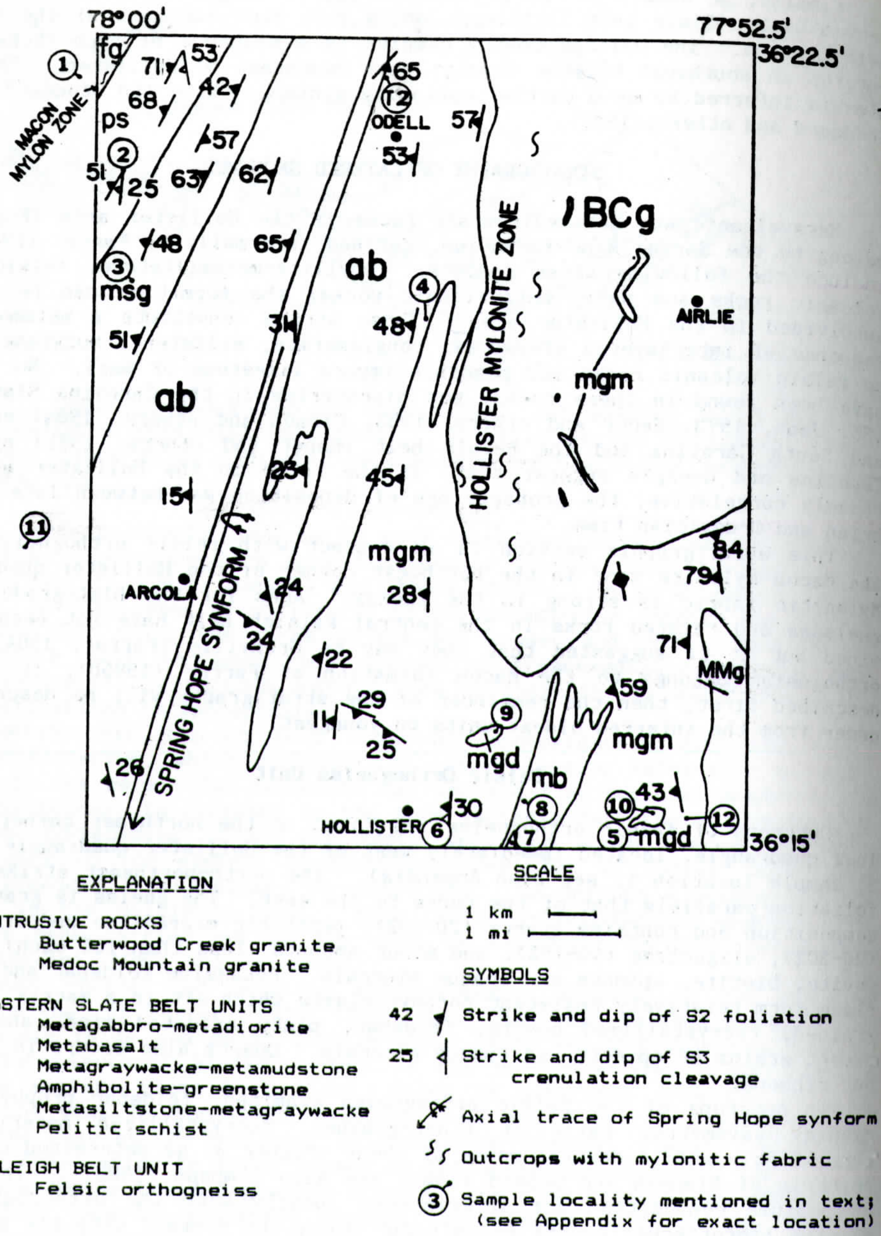

Figure 2. Generalized bedrock geologic map of the Hollister quadrangle (post-Paleozoic rocks are not shown). Age relationship between units of the Eastern Slate belt terrane and felsic orthogneiss of the Raleigh belt terrane is unknown.

Appendix). Pelitic schist is typically gray when fresh and many samples contain abundant (up to 20%) prophyroblasts of garnet and sparse (less than 5%) staurolite. Other, more quartzose schists are devoid of prophyroblasts and are comprised of quartz (10 to 50%) and white mica (20 to 70%), with lesser amounts of chlorite and opaque minerals.

Figure 3. Photomicrograph, under plane polarized light, of K-feldspar porphyroclast in felsic orthogneiss unit (in Macon mylonite zone of Farrar, 1985a). Asymmetric tails of recrystallized grains, at upper right and lower left, suggest dextral sense of shear, applying the criteria of Simpson and Schmid (1983), and Simpson (1986). Sample location is number 1 on Figure 2. For this and other sample locations, consult Appendix. View is down-dip, toward the northwest; foliation strikes N15E and dips 53 NW. Scale bar is 0.5 mm.

Metasiltstone-Metagraywacke Unit

This unit contains primarily laminated, fine-grained, quartz-rich rocks that are pinkish-gray in fresh hand sample but weather to a dull yellow-green. Quartz (50-90%), sodic plagioclase (5-30%) and white mica (5-15%) are phases common to all of the metasiltstones. Additional minor phases which may be present include biotite, chlorite, epidote, clinozoisite, garnet and opaque minerals. Interbedded with these rocks in subordinate amounts are metagraywacke, gray-green, siliceous metamudstone, massive greenstone, hornblende schist and muscovite schist. The layering is exposed discontinuously in a roadcut (Figure 2, location 3; see Appendix).

Microscopic examination of the metasiltstone reveals that laminae, commonly less than one mm thick, are characterized by variations in quartz grain size. Although recrystallization textures are evident among quartz grains throughout the rocks, individual laminae are laterally continuous and are traceable at the outcrop scale. It is therefore concluded that, although metamorphic textural adjustments have probably slightly altered original grain size, the laminations are the result of sedimentary processes as opposed to metamorphic differentiation. Sheet silicates typically are distributed randomly throughout the rock and are only locally concentrated along individual laminae. A weak fissility is developed parallel to bedding planes in metasiltstones. This foliation is observed as a pervasive schistosity in the interbedded schists. The strike of bedding and schistosity is approximately N20E with dips ranging from 30 to 65 degrees west.

A pyroclastic origin is ascribed to part of the unit. Evidence for volcanic contribution includes the occurrence of euhedral to subhedral plagioclase laths and subrounded quartz clasts suspended within the much finer grained matrix. The recrystallized nature of the matrix of these rocks makes it impossible to determine whether the protoliths were epiclastic siltstones contaminated by an influx of volcanic material contemporaneous with deposition, or tuffs subsequently reworked by water currents. The metasiltstone grades westward into the pelitic unit; micaceous quartzite and schist are interbedded with the siltstone along the transitional boundary. Similarly, the metasiltstone is interlayered with chlorite and hornblende schist as the contact with the overlying mafic unit to the east is approached.

Amphibolite Unit

This unit is a sequence of mafic lithologic types including dark greenish black amphibolite, hornblende schist and epidote-rich greenstone (interbedded amphibolite and greenstone are exposed along a roadcut at location 4 in Figure 2; see Appendix). The unit is stratigraphically overlain by a sequence of undifferentiated flyschoid metasedimentary rocks.

Rocks of the amphibolite unit show a range in textures from massive epidote greenstone containing quartz (15-50%), epidote (15-50%) and sparse actinolite (0-15%), to variously foliated or lineated amphibolite with blue-green hornblende (30-70%), plagioclase (20-40%), and epidote (5-20%), with or without quartz (up to 15%). Common minor constituents are titanite, ilmenite, rutile, chlorite and brownish yellow or brown biotite. Amygdules are preserved in some greenstone and amphibolite and suggest that many of these rocks are metabasalt or metadiabase; amygdules contain quartz with occasional epidote-group minerals. If the epidote greenstones were originally igneous rocks, they must have undergone chemical alteration prior to or during metamorphism, inasmuch as they are too Ca, Fe, and Si-rich to represent any common igneous rock chemistry. The significant mineralogical differences between the greenstones and the amphibolites are variations in modal abundances of epidote and amphibole, which may be attributed to progressive chemical modification of original basalts.

Textural variations among the amphibolites are best observed microscopically. One striking aspect of these rocks is the untwinned nature of the plagioclase in most samples. Twinned plagioclase does occur, albeit rarely, and the random orientation displayed by the tabular grains is suggestive of relict igneous texture. Blue-green hornblende is most often poikiloblastic and occurs as coarse prisms or as radiating, sheaf-like aggregates. The greater the alignment of hornblende, the more prominent is the foliation in the rock. Hornblendes in amphibolite interbedded with metasedimentary rocks at the base of the unit are generally finer grained, relatively inclusion-free, slender prismatic crystals that are well aligned and impart a nematoblastic fabric to the rock. With the exception of amygdaloidal samples, most amphibolites are too thoroughly recrystallized to retain primary features which might be of use in determining their protoliths.

Metagraywacke-Metamudstone Unit

This unit comprises a metamorphosed package of interbedded graywacke, mudstone (including argillite, phyllite and schist), siltstone and conglomerate with intermediate to felsic volcanic rocks. The unit is overlain by metabasalt in the chlorite zone of regional metamorphism and overlies amphibolite at higher metamorphic grade. The multiply interlayered sequence is traceable from chlorite grade rocks with obvious sedimentary protoliths up through schists and feldspathic quartzites at almandine garnet grade in which relict sedimentary textures are nearly totally destroyed. The lithologies present, the nature and scale of their interlayering, and the discernible primary features all suggest that this unit represents a series of metamorphosed sediments deposited as subaqueous turbidity flows (see Walker, 1984).

Chlorite and biotite grade graywackes contain subrounded to angular fragments of quartz, sodic plagioclase and rock fragments in a matrix of white mica, chlorite and opaque minerals. Epidote is an occasional constituent. Rock fragments consist of polycrystalline quartz, plagioclase glomerocrysts and chloritic fragments. Chloritic fragments with boundaries that are partially remolded to conformity along contacts with more coherent grains are probably clay rip-up clasts (Figure 4A; location 5 on Figure 2) torn from underlying mudstones by high-density gravity flows which deposited these impure sands (*cf.* Sigurdsson and others, 1980). Other chloritic fragments behaved more coherently and may be altered mafic volcanic fragments. Polycrystalline quartz grains may be fragments of gneiss, hydrothermal quartz veins or quartzite. Plagioclase glomerocrysts are, like monocrystalline plagioclase grains, derived from a volcanic source area based on the subhedral, lath-like form exhibited by most of the grains. Nearly all plagioclase shows some sericitization but very few of the grains are significantly rounded. A number of the quartz grains display relict beta-quartz forms and are embayed, further attesting to the volcanogenic nature of the graywackes.

Metagraywacke is regularly interlayered with metamudstone or phyllite and local thin metasiltstone or metaconglomerate beds. The smallest sandstone beds observed are approximately three cm thick; thickness of the largest beds is unknown due to lack of outcrop, but exceeds six meters. Load casts occur along the bases of some of the graywacke beds where the denser sands have sunk into underlying mud. Weakly recognizable graded bedding is observed locally.

Mudstone in the low-grade rocks is comprised of the mineral assemblage quartz, sodic plagioclase, white mica, chlorite (or biotite) and opaque minerals. Most of the mudstones are non-fissile, blocky rocks which may or may not be laminated and which sport a weak spaced cleavage. Others are well-foliated phyllites which contain a larger proportion of platy minerals (chlorite, white mica). These argillaceous rocks occur in thin laminae less than one mm thick as well as in beds that are meters or tens of meters thick. Mudstone is represented by coarse-grained micaceous schist in the highest grade portion of the unit.

Basal conglomerates occur at several stratigraphic levels within the unit, each grading upward into graywacke (for example, location 6, Figure 2). The top of the graded bed is defined by a sharp contact with an overlying basal conglomerate of a succeeding graded bed. A local conglomeratic facies may be defined within the metagraywacke-metamudstone unit southeast of the town of Hollister. Conglomerates display some variations in hand sample but microscopic examination reveals that they are compositionally and texturally similar. The rocks are largely clast- supported and the small amount of distinguishable matrix may be classified as graywacke sandstone. Fragments are coarse sand, granules and pebbles of mostly fine-grained, impure sandstone or siltstone. Included also are rounded pebbles of highly strained polycrystalline quartz, and subrounded to angular pebbles of mafic to felsic volcanic rocks and argillaceous rocks.

Thin lenses or beds of felsic to intermediate volcanic material are intercalated with the volcaniclastic sediments, but are untraceable for any distance. Some volcanic rocks are silicified and superficially resemble chert layers; upon microscopic inspection, however, scattered crystals of plagioclase are observed suspended in a matrix of microcrystalline quartz and sericite. Another type of volcanic rock (Figure 2, location 7) is a spherulitic metafelsite containing abundant euhedral paramorphs of beta quartz (Figure 4B). The spherulites (Figure 4C) are intergrowths of sodic plagioclase with quartz. A mafic or intermediate metatuff has also been found in this unit, but has been extremely altered so that no relict volcanic textures remain. It has the assemblage chlorite, albite, epidote, calcite and white mica.

Evidence for flyschoid or turbidite sedimentation of the metagraywacke-metamudstone unit is abundant; besides the predominance of argillaceous sediments and of graywacke as the only sandstone type, many other characteristics of flysch deposits as listed by Pettijohn (1975, p. 578) are common to the unit. These include the clastic nature of the sediments, their close associ-

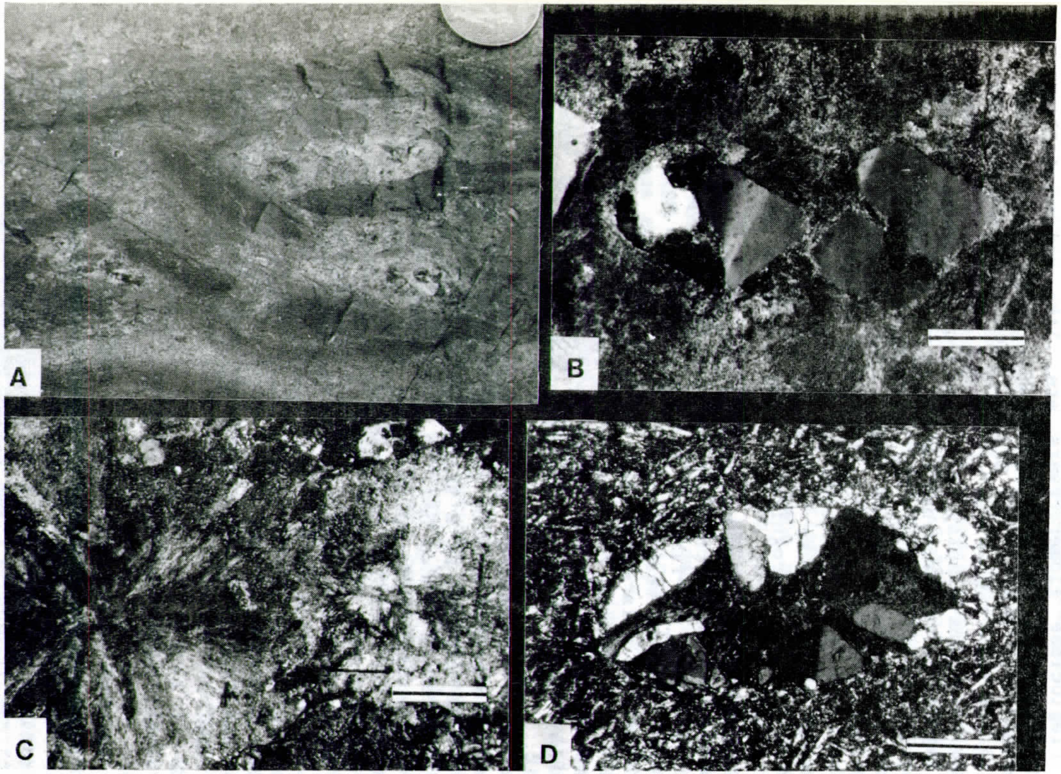

Figure 4. Selected petrographic features. B, C, and D are photomicrographs taken under crossed polars; scale bars in B, C, and D each represent 0.5 mm. A. Elongate clay rip-up clasts in boulder of metagraywacke-metamudstone unit. The boulder is not in place. Note (partial) dime for scale. Sample location 5, Figure 2. B. Beta- quartz paramorphs in felsic volcanic facies of the metagraywacke-metamudstone unit. Sample location 7, Figure 2. C. Spherulites, consisting of intergrown quartz and sodic plagioclase, from metafelsite of metagraywacke- metamudstone unit. Sample location 7, Figure 2. D. Amygdaloidal metabasalt (from metabasalt unit) showing elongate amygdule filled with epidote, chlorite and calcite in fine-grained relict basaltic matrix. Sample location 8, Figure 2.

iation with metabasalts of oceanic affinity (Kite and Stoddard, 1984), occurrence of graded bedding with occasional ripple laminations, absence of large-scale cross-bedding, the uniform and somewhat rhythmic nature of bedding, and range in thickness of graded graywacke beds from a few cm up to a few meters. The occurrence of coarser grained pebbly sandstone and conglomerate may be ascribed to deposition in the upper portions of submarine debris fans. The association of the coarse intervals with shale-sandstone turbidite sequences is typical of such facies (Walker, 1984).

Metabasalt Unit

A distinctive unit of metabasalt occurs in the southeastern part of the Hollister quadrangle. Good exposures of the unit are found along a two-km stretch of a creek just east of the town of Hollister (Figure 2, location 8).

Most of the rocks comprising the unit are amygdaloidal basalts which have been metamorphosed to chlorite grade. In hand sample, the metabasalts are fairly massive and gray-green to greenish black. The groundmass mineral assemblage is albite (35-50%), quartz (5-20%), chlorite (10-30%), epidote (5-20%) and locally calcite (up to 15%), and/or minor titanite, apatite and opaque minerals. Sericitization and saussuritization of plagioclase implying its conversion from a former, more calcic form are common. Titanite and opaque oxide, probably ilmenite, are partially altered to brownish patches of leucoxene. Amygdules typically consist of one or more of the following: quartz, chlorite, calcite and epidote (Figure 4D). Microphenocrysts and glomerocrysts of plagioclase occur in some samples as do epidote grains that are larger than the typical matrix grains. Relict mafic igneous phases (e.g., olivine and pyroxene) have not been observed either in the groundmass or as phenocrysts in any of the rocks.

Relict igneous texture is recognized by the abundant randomly oriented, subhedral laths of plagioclase in the groundmass. Some samples, however, are significantly finer grained and primary textures are not observable even under high magnification. Others are not amygdaloidal and may or may not display the random plagioclase orientation. Still, these rocks are all related by similar mineralogy and plagioclase composition; the maximum anorthite content as determined by the Michel-Levy method is An 08, indicating that, although primary grain shapes were retained, the plagioclase composition has been modified during metamorphism.

Hyaloclastite occurs within the unit near the contact with the underlying flyschoid metasediments. The brecciated texture of this rock was apparently developed during interaction of hot basaltic lava with water and water-rich sediments at the margins of a flow during subaqueous extrusion (Williams and others, 1982, p. 271). Fragments are angular and range in size from one mm to seven cm or larger. The fragments are equant to lensoidal and some are flattened fiamme. Others are vesicular and consist of opaque material. Microscopically, the matrix of the rock appears to be somewhat gradational into the fragments, although the boundaries are well defined by matrix-fragment compositional differences.

Seven specimens of metabasalt and amphibolite from the Hollister area were analyzed for major oxides and several trace elements by X-ray fluorescence spectrometry. Two representative analyses, one for each rock type, are shown in Table 1, along with analyses of two meta-intrusive rocks (see below). Chemically, the Hollister metabasalts and amphibolites show evidence of significant alteration, with large ranges in the more mobile oxides MgO (3.6-11.7 weight percent), CaO (4.7-12.4), and Na_2O (0.4-3.3). However, in terms of the more stable elements Ti, Y, and Zr, the analyses cluster more tightly. On a Ti-Y-Zr discrimination plot (Figure 5), they fall in field B, a region of chemical overlap among ocean floor basalts, calc- alkali basalts, and low-potassium tholeiites (Pearce and Cann, 1973). For comparison, 15 of 18 analyses of metabasalt and amphibolite from the nearby Halifax County complex (Kite and Stoddard, 1984) plot in field A, which is restricted to low-potassium tholeiites. The chemical evidence is consistent with the suggestion that the Hollister mafic rocks belong to the same low-potassium tholeiitic suite as the Halifax complex rocks. Inasmuch as they interfinger with metasedimentary rocks, the Hollister metavolcanics may occupy a higher stratigraphic or more distal position relative to the volcanic source.

INTRUSIVE ROCKS

Metagabbro-Metadiorite

A mafic to intermediate plutonic complex (mgd in Figure 2) intruded to the upper levels of the Eastern Slate belt stratigraphic sequence prior to regional metamorphism. These intrusive rocks occur in two separate areas within the chlorite zone in the southeastern corner of the Hollister quadrangle (Figure 2, sample locations 9 and 10; see Appendix).

The intrusive rocks are massive with color index of 40 or greater and contain albite (35-50%), quartz (5-15%), chlorite (5-20%), epidote (5-20%), calcite (0-10%), and minor amounts (less than 5% each) of apatite and a nearly opaque alteration product pseudomorphous after original clinopyroxene (Figure 6). These pseudomorphs preserve original exsolution textures of the pyroxene. Major element X-ray fluorescence analyses on two specimens from the complex (Table 1) reflect the range of compositions present. Sample HRB 237 (Figure 2, location 9) is a metagabbro and HRB 256 (Figure 2, location 10) is a metadiorite. Major oxides (Table 1) are very similar to analyses of plutonic rocks from the Halifax County complex (Kite and Stoddard, 1984), suggesting that the Hollister plutonic rocks as well as the volcanics, are closely related to their Halifax counterparts.

Granitoid Intrusives

The metamorphosed volcaniclastic sequence is intruded by two late Paleozoic granitoid bodies emplaced subsequent to the regional metamorphic peak. Approximately one-third of the map area is underlain by the massive to strongly foliated, compositionally variable granitic rocks. The southwest corner of the 292 ± 30 Ma. Butterwood Creek pluton (Farrar and others, 1981; Russell and others, 1985) comprises the northeast portion of the Hollister quadrangle. A small granite stock, the Medoc Mountain pluton, occurs in the southeastern corner of the area and has been dated at 301 ± 6 Ma. (Fullagar and Butler, 1979). Outcrops of this latter intrusion are scarce. A study of sulfide mineralization associated with the intrusion (Harvey, 1974) includes descriptions of several drill cores which reveal the occurrence of massive, medium-grained biotite granite at shallow depth. The north-south linear ridges which topographically define the location of the pluton are held up by a sheet of resistant vein quartz, some of it mineralized, which apparently forms a thin cap on the pluton.

The southwestern corner of the Butterwood Creek pluton, represented by porphyritic to medium grained biotite granite (IUGS classification; see Streckeisen, 1981) and leucogranite (color index less than 10), occupies the northeastern third of the Hollister quadrangle. The petrography, field relationships, and chemistry of these rocks are under investigation (Stoddard and others, unpublished data).

Upon intrusion of the pluton, some of the overlying country rocks were engulfed by the granitic magma, perhaps by the process of magmatic stoping. Erosion has left isolated screens of host rock surrounded by the granite. A few of these screens are traceable along strike for over one km. They are mostly pelitic schist which show the effects of contact metamorphism (see METAMORPHISM) and are presumably equivalents of the graywacke-mudstone facies metasedimentary rocks that occur roughly along strike to the south. A few screens consist of amphibolite or calc-hornfels which are probably thermally metamorphosed basaltic rocks and impure limestones or marls, respectively.

Table 1. Chemical compositions of selected meta-igneous rocks from the Hollister area.

Specimen	HRB-260	HES-K2	HRB-237	HRB-256
Rock type	meta-basalt	amphibolite	meta-gabbro	meta-diorite
		oxides (weight %)		
SiO_2	50.4	53.0	48.9	55.8
TiO_2	1.16	1.22	0.61	0.79
Al_2O_3	15.0	14.8	15.8	14.4
Fe_2O_3*	14.3	12.6	8.7	13.4
MgO	9.4	4.7	9.0	4.4
CaO	4.66	9.68	13.06	5.02
MnO	0.31	0.21	0.19	0.24
Na_2O	2.8	1.8	2.2	2.7
K_2O	0.26	0.34	0.19	0.58
P_2O_5	0.10	0.21	0.08	0.32
L. I.	n.d.	1.4	n.d.	1.7
Total	98.39	99.96	98.73	99.35
		Trace elements (ppm)		
Y	30	34	n.d.	n.d.
Zr	64	78	n.d.	n.d.

*total Fe as Fe_2O_3
L.I.: weight loss on ignition
n.d.: not determined
For details concerning analytical procedures, the interested reader may consult Boltin (1985)
See Appendix for sample locations

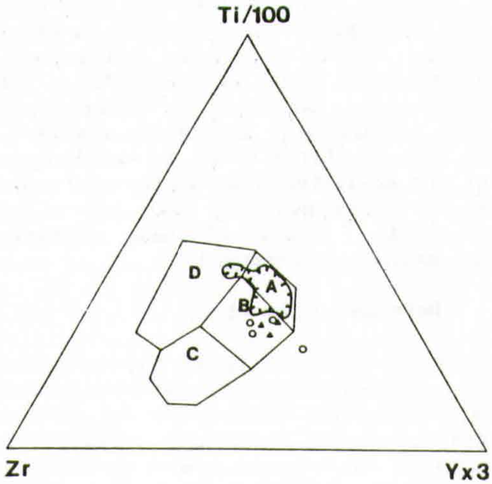

Figure 5. Ti-Zr-Y discrimination diagram for Hollister area metabasalts (circles) and amphibolites (triangles). Fields (from Pearce and Cann, 1973) are as follows: low-potassium tholeiites, A and B; ocean floor basalts, b; calc-alkali basalts, B and C; within-plate basalts, D. Analyses of 18 metabasalts and amphibolites from the Halifax County complex (Kite and Stoddard, 1984) fall in the area indicated by the hachures.

The western border of the Butterwood Creek pluton has undergone intense late Paleozoic deformation and a strong mylonitic fabric is developed there. Intensity of mylonitic deformation decreases eastward toward the pluton's interior.

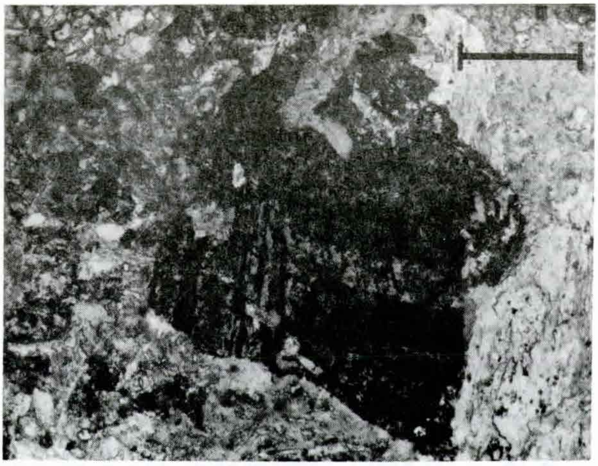

Figure 6. Pseudomorph, largely composed of chlorite and opaque minerals, after clinopyroxene in metagabbro unit. The opaque minerals appear to preserve original pyroxene exsolution or cleavage planes. Plane-polarized light; scale bar 0.5 mm. Sample location 9, Figure 2.

STRUCTURE

Rocks in the Hollister area exhibit two, or possibly three, tectonic foliations in addition to primary bedding.

Primary Bedding S_0

Primary bedding surfaces, S_0, are conspicuous on scales from less than one mm up to the map-scale layering which defines the various units. Bedding is especially obvious in the low-grade metasedimentary rocks of the metagraywacke-metamudstone unit, where delicate primary sedimentary features are locally visible (see discussion above under *Stratigraphy*). Outcrops containing primary evidence of stratigraphic younging direction include graded conglomerate at Hollister, where bedding dips moderately eastward and is upright (Figure 2, location 6), and along Gunter's Creek in the Inez quadrangle, just west of the western edge of the Hollister quadrangle, where graded bedding in metasiltstone and metagraywacke suggests that the west- dipping beds are overturned (Figure 2, location 11). In both of these locations, the dominant regional foliation (S_2) dips moderately westward.

Deformation Event D_1

A possible separate early foliation, S_1?, is observed microscopically in metapelites in the highest grade portion of the study area. This foliation is transposed by a regional schistosity (presumably S_2) and is observed only locally where it is preserved outlining the hinges of microfolds. S_1 may be the same as S_0 but no conclusive evidence of compositional layering has been observed in S_1. If S_1 is a deformational fabric independent of S_0, its tectonic significance has been lost in overprinting relationships related to subsequent Paleozoic events.

Deformation Event D_2

The dominant foliation observed is manifested as a pervasive schistosity in the high-grade northwest portion of the area and becomes less intense in the lower grade southeast corner of the quadrangle. This correlation between the pervasiveness of S_2 and metamorphic grade suggests that the formation of

S_2 was syn-metamorphic. The prominent S_2 foliation parallels compositional layering, S_0, in most places, suggesting the presence of (unmapped) isoclinal F_2 folds. Several outcrops in which S_2 cross-cuts layering obliquely may lie near the hinge zones of such folds. In the low- grade southeastern corner of the quadrangle, S_0 was observed to be transected by a later foliation in exposures of metamudstone (Figure 2, Location 12).

The lithologic discontinuity separating the felsic orthogneiss in the northwest from the sequence of metasedimentary and metavolcanic rocks is oriented parallel to S_2; indeed, the mylonitic foliation in the felsic orthogeniss, also interpreted as a manifestation of S_2, corresponds with the regional schistosity in the stratified rocks to the east, and both were affected by F_3 folding. Farrar (1985a) considered development of the mylonitic fabric along this zone to have been contemporaneous with development of the D_3 Hollister mylonite zone and he includes the felsic orthogneiss in a zone of ductile deformation which he calls the Macon mylonite zone. However, this study suggests that juxtaposition of the gneissic terrane and the volcanogenic sequence, and formation of the mylonitic fabric in the felsic orthogneiss (and Macon zone), occurred prior to the D_3 event, either before or during the D_2 event. This interpretation does not preclude the possibility of subsequent reactivation along the Macon zone during the D_3 event.

Deformation Event D_3

Deformation event D_3 is responsible for the large-scale structures observable in the Hollister area. D_3 is interpreted to be an Alleghanian event, due to its effects on the Alleghanian Butterwood Creek pluton (see below). D_3 produced abundant microscopic to macroscopic folds, the larger of which are defined by changing orientation of strike and dip of S_2 (and apparently S_0) and by the associated map pattern where contacts are folded.

Field evidence indicates that intrusion of the Butterwood Creek pluton preceded, or was at least partially synchronous with, the D_3 event. D_3 resulted in development of the ductile Hollister mylonite zone (see Farrar, 1985a) which deforms the western margin of the intrusion. Varying degrees of intensity of D_3 are observable in rocks of the Butterwood Creek pluton within the mylonite zone. Effects range from zones of porphyroclastic mylonite to zones of incipient mylonitization in which augen gneiss occurs, to zones of granitic orthogneiss and lineated granite. South of the Hollister quadrangle, mylonitic rocks attributed to the Hollister zone have been mapped by Farrar (1985a) and by Kite and Stoddard (1984). However, in the Hollister quadrangle, observed mylonitic effects were limited to the granitoid rocks of the Butterwood Creek pluton. Unfortunately, an apparent lack of outcrop in the key area south of the pluton does not allow mapping of the mylonite zone out of the pluton and into the country rock.

In this study, the mylonitic foliation of the Hollister zone, with an average strike of N05 to 10^{o}E and a steep westward dip, is classified as S_{3m}. In addition, S_3 foliation is observed locally as a crenulation cleavage (S_{3c}) oblique to S_2 foliation in metapelites and elsewhere may parallel S_0 and S_2 along the limbs of F_3 folds. As it is unknown whether in fact the crenulation cleavage is precisely equivalent to the mylonitic foliation, they are distinguished by subscripts (c and m). In the Hollister zone, planar and linear fabrics attributed to D_3 cluster tightly, whereas in the schist they are more diffuse (Figure 7).

Figure 8 displays a contoured equal-area projection of poles to S_2 along with D_3 linear fabric elements. These include four measured lineations as well as lines defined by the intersection of the S_2 schistosity with the S_{3c} crenulation cleavage in outcrops of metapelites. The linear features statistically plunge gently toward the south-southwest. The contoured poles display a strong maximum, with a suggestion of some spread toward the northwest about a great circle. All the data of Figure 8 are consistent with the existence of a nearly isoclinal, gently southwestward-plunging post- D_2 fold with

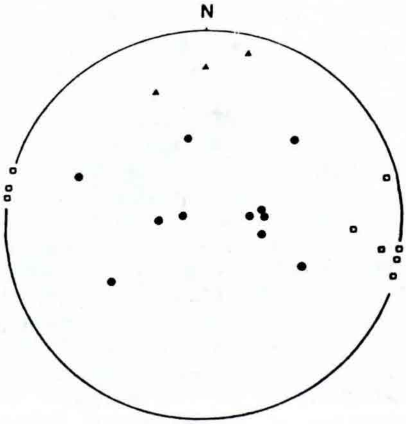

Figure 7. Lower-hemisphere equal-area projection showing: (1) open squares - six poles to S_{3m} foliations in outcrops of the Butterwood Creek pluton from within the Hollister mylonite zone; (2) triangles - three biotite mineral lineations from some of the same outcrops; and (3) dots - eleven poles to S_{3c} crenulation cleavage from pelitic schist west of the Hollister zone.

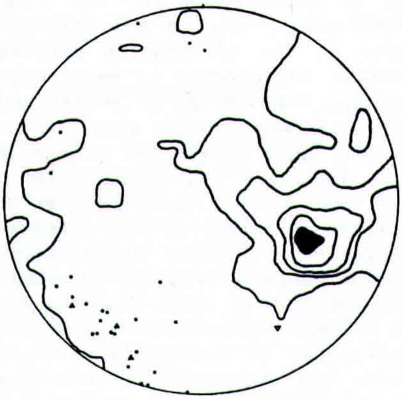

Figure 8. Contoured lower-hemisphere equal-area projection of poles to 115 S_2 foliations. The dominant westward dips are due to overturning and nearly isoclinal form of the F_3 Spring Hope synform and parasitic folds on its east limb. Contours are 14, 10, 6, 3, and 0.5% per 1% area. Individual points plotted include three measured crenulation lineations (solid triangles), one measured mineral lineation (open triangle), and 25 intersections determined for S_2 and S_{3c} measured in outcrops of pelitic schist (solid dots). These linear features suggest a gentle south-southwest plunge for the Spring Hope synform.

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an axial surface that dips moderately to steeply northwestward. This fold corresponds well with the (F_3) Spring Hope synform of Farrar (1985a); the outcrop pattern in the Hollister quadrangle suggests the occurrence of a number of smaller parasitic folds along its east limb (Figure 2). Although the data are sparse, primary structures indicative of stratigraphic younging directions, outlined above, indicate that the Spring Hope synform is in fact a syncline overturned to the east, provided that S_2 and S_0 are parallel. In the northwestern Hollister quadrangle, outcrops of pelitic schist located within the western limb of the Spring Hope synform (for example, sample location 2, Figure 2) display S_{3c} crenulation cleavage that consistently dips west at shallower angles than the S_2 schistosity, lending further support to the inference that the fold is overturned.

METAMORPHISM

The volcanic-sedimentary sequence and gabbro-diorite intrusive of the Hollister area have undergone at least one episode of prograde regional metamorphism and a single retrograde event. Contact metamorphic assemblages are superposed on regional assemblages in a thermal aureole around the late Paleozoic Butterwood Creek pluton as well as in screens of metasedimentary rock within the pluton.

A Barrovian sequence of chlorite, biotite, garnet and staurolite zones is mapped in the volcanogenic sequence (Figure 9) and defines a steep metamorphic gradient. Distance from the biotite-in isograd of metagraywackes and metapelites to the first occurrence of staurolite in metapelites is approximately eight km across the present erosion surface. The metamorphic grade of the Raleigh belt orthogneiss to the west is unknown. However, sillimanite, kyanite and andalusite have all been reported a few km to the west of the study area (Farrar, 1985a; Stoddard and McDaniel, 1979; E. F. Stoddard, unpublished data).

Regional Prograde Event

Regional metamorphism was in part synchronous with D_2 deformation but the thermal peak of this event was not achieved until final development of the S_2 axial planar schistosity. Metamorphic muscovite, biotite and chlorite define the S_2 foliation in metasedimentary rocks. The idioblastic form of garnet and staurolite, and weak deflection of S_2 about garnet suggest that growth of porphyroblasts continued beyond the climax of S_2 formation; the prograde event culminated with the crystallization of post-D_2 helicitic (post-tectonic) staurolite porphyroblasts that enclose opaque inclusions aligned parallel to S_2 (Figure 10). Randomly oriented coarse muscovite in highest grade pelitic schists also grew post-tectonically with respect to the pervasive S_2 schistosity. Poikiloblastic muscovite in garnet- grade metagraywacke and metasiltstone is probably of equivalent generation to the coarse muscovite in metapelites. Skeletal ("fish-net") garnets, also indicative of post-tectonic crystallization (Spry, 1969, p. 271), are prevalent in semipelitic rocks within the garnet zone. Age relationships between the prograde event and D_3 are not clear owing to the lack of development of a pervasive S_3 foliation. The isograd map (Figure 9) offers no evidence for F_3 succeeding regional metamorphism, but control on the orientation of the isograds is tenuous as denoted by the dashed lines.

Albite, chlorite and epidote are stable phases in mafic rocks at low grade; calcite is a common constituent in the metabasites. Actinolite is absent at low grade, probably due to high XCO_2 (Billings and White, 1950). Blue-green hornblende first appears in amphibolite in the middle to upper biotite zone of associated metasedimentary rocks. Actinolite occurs in greater abundance above the garnet-in isograd than below, presumably due to elimination of CO_2 by decarbonation reactions at the higher metamorphic grade. It is subordinate to hornblende, and typically coexists with hornblende as a minor phase in epidote-rich greenstones. Rarely is actinolite observed as the sole amphibole in the greenstones though hornblende quite commonly occurs alone in these rocks. Actinolite apparently persists to the highest grade of metamorphism in the area.

The anorthite component of plagioclase in metabasites increases abruptly to calcic oligoclase, generally corresponding with the first appearance of hornblende. However, it has not been determined whether the hornblende and oligoclase isograds correspond precisely, due to a lack of rocks of appropriate composition in the vicinity of the biotite-in isograd of associated metapelites.

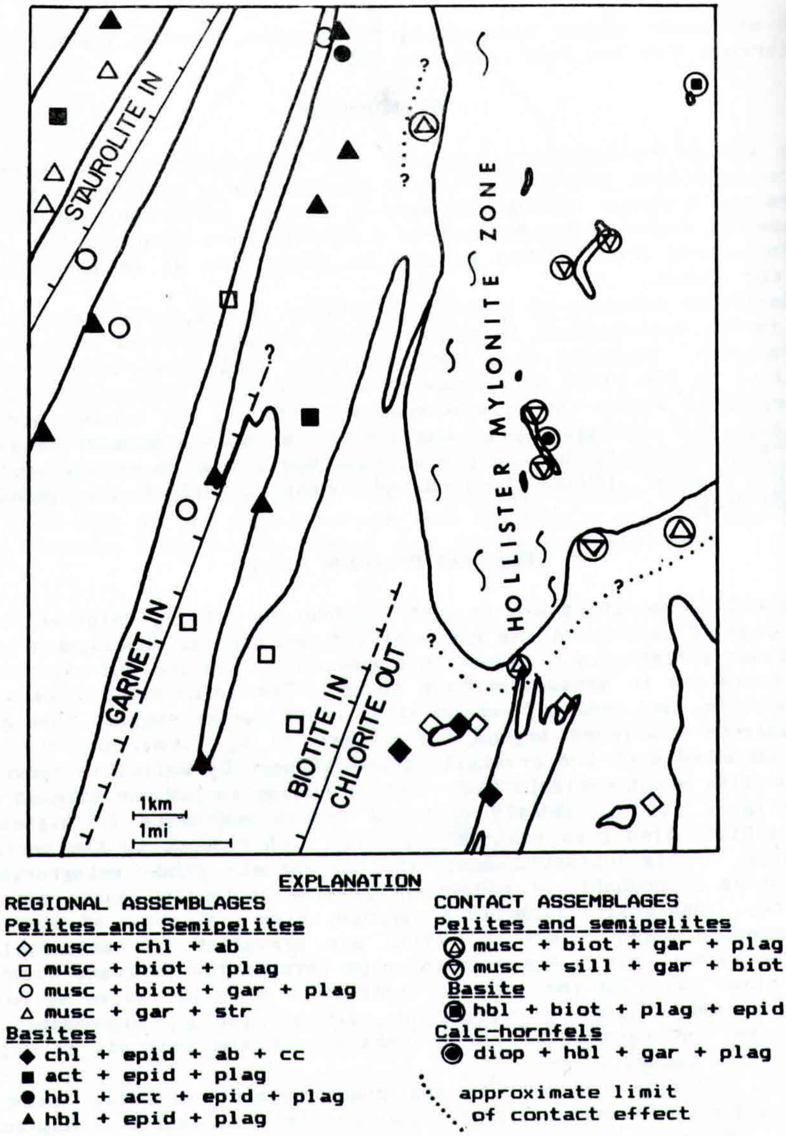

Figure 9. Metamorphic mineral assemblages and regional isograds in the Hollister area. Open symbols are pelitic rocks; closed symbols are mafic. Single symbols represent assemblages interpreted to be the result of prograde regional metamorphism; contact metamorphic overprint related to the intrusion of the Butterwood Creek pluton is indicated by the encircled symbols. Mineral abbreviations: musc, white mica; chl, chlorite; ab, albite; biot, biotite; plag, oligoclase-andesine; gar, garnet; str, staurolite; epid, epidote-clinozoisite; cc, calcite; act, actinolite; hbl, blue-green amphibole; sill, sillimanite; diop, diopside.

Contact Metamorphism

Contact metamorphism accompanied intrusion of the late Paleozoic Butterwood Creek pluton. Contact effects are recognizable as a relatively thin aureole rimming at least the southern end of the pluton and as screens of

Figure 10. Staurolite porphyroblasts in muscovite-biotite schist helicitically enclosing opaque grains aligned parallel to S_2 schistosity of the matrix, suggesting post-D_2 growth of at least some staurolite. Plane-polarized light; scale bar 0.5 mm. Sample location 2, Figure 2.

pelitic and mafic material engulfed by the pluton during intrusion. Pelitic and semipelitic rocks from a small area within the contact aureole south of the intrusion (Figure 9), contain coarse muscovite poikiloblasts and locally biotite and pinhead almandine(?) garnets in rocks regionally metamorphosed only to chlorite grade. Contact and regional metamorphic effects are generally indistinguishable in the higher grade rocks along the western border of the pluton. Screens of country rock within the pluton consist of pelitic, mafic and calc hornfels. Mineral assemblages of these rocks are typical of hornblende hornfels facies metamorphism (Figure 9). Pelites contain quartz and muscovite and one or more of the following: garnet, sillimanite, biotite, plagioclase, chlorite (retrograde), tourmaline and/or opaque minerals. Calc hornfels contains blue-green or brown-green hornblende, andesine or labradorite, quartz, garnet, titanite, opaque minerals, diopside and/or epidote group minerals.

Muscovite grown during the thermal event is porphyroblastic and often is observed growing across a pre-existing foliation at high angles. Sillimanite occurs mainly as bundles of fibrolite mimetically following the earlier foliation and enclosed in late muscovite porphyroblasts. Needlelike sillimanite is locally found as inclusions in quartz grains.

Retrograde Event

A late retrograde event is believed contemporaneous with D_3 mylonitization. Strongest evidence for retrogression is found in the relatively high grade pelitic rocks of the slate belt sequence located in the northwest part of the Hollister quadrangle and in thermally metamorphosed pelitic rocks in screens within the Butterwood Creek pluton. Replacement textures include chlorite forming partial reaction rims on garnet and staurolite, chlorite replacing biotite, and muscovite pseudomorphs after fibrolite which grew during the contact metamorphic event.

CONCLUSIONS

Belt Definitions

Because in many cases the regional metamorphism is a late event superimposed upon more fundamental lithotectonic terrane boundaries, the belt concept is only partially useful in constructing viable tectonic models, especially in the Piedmont. In the Hollister area, amphibolite-facies rocks (*metamorphic* Raleigh belt) and greenschist-facies rocks (*metamorphic* Eastern Slate belt) all belong to a single stratigraphic package of interbedded volcanic rocks and volcanogenic flyschoid sediments. The steep metamorphic gradient imposed on the sequence is a late Paleozoic feature (Russell and others, 1985). In terms of tectonic models, the more fundamental lithotectonic ("terrane") boundary is west of the metamorphic front, where this volcanogenic sequence of apparent oceanic affinity is in contact with felsic orthogneiss of probable continental affinity (see also Farrar, 1985a,b).

Timing of Deformation and Metamorphism

Detailed mapping in the Hollister quadrangle suggests that the boundary between the felsic orthogneiss (Raleigh belt *terrane*) and the variably metamorphosed volcanogenic sequence (Eastern Slate belt *terrane*) is a D_2 (or earlier) fault. Recent models (Wylie, 1984; Farrar, 1985a) suggest that a volcanogenic terrane (including both the Eastern and Carolina Slate belts) was thrust over a continental (Raleigh belt) terrane, which is now exposed by erosion as a tectonic window through the overlying thrust sheet. Farrar (1985a) considers rocks of the Raleigh belt terrane to likely be part of the North American craton and Grenville in age. He suggests the decollement was overturned as a result of major D_3 folding which resulted in formation of the Macon and Hollister mylonite zones along attenuated limbs of map-scale folds. An alternative interpretation is that the continental terrane (felsic orthogneiss) was thrust over the volcanogenic sequence. Yet another interpretation might be that the terrane boundaries are strike-slip faults and that thrusting is not involved. None of these models can be excluded on the basis of mapping and structural data now available. Under Farrar's (1985a) interpretation, the mylonitic foliation with its indicators of dextral sense of shear in the felsic orthogneiss of the Macon zone would have resulted from reactivation along the decollement. Evidence presented above, however, suggests that the mylonitic fabric in the felsic orthogneiss was deformed by F_3 folding, and therefore the mylonitic fabric of the Macon zone is pre-D_3, and probably a D_2 feature. Deformation of the post-regional metamorphic, 292-Ma Butterwood creek pluton (Russell and others, 1985) in the D_3 Hollister mylonite zone, and deformation of regional S_2 foliation attributed to intrusion of the pluton, constrain the age of D_3 to late Paleozoic time (Alleghanian orogeny) and of D_2 to pre- or syn-intrusion. Petrographic evidence (see METAMORPHISM) that D_2 was in part synchronous with regional metamorphism documented as late Paleozoic (Russell and others, 1985), and the increasing pervasiveness of S_2 with increasing metamorphic grade, suggest that D_2 is also Alleghanian. If an earlier D_1 event occurred (the evidence is admittedly weak), it may have been pre-Alleghanian, but available information cannot constrain it further.

Tectonic Implications

The component of strike-slip motion along the Macon zone implied by the sense of shear in feldspar porphyroclasts of the felsic orthogneiss unit may be explained by dextral movement along the zone either contemporaneous with D_2 thrusting, or during D_3 reactivation along an earlier (D_2?) decollement. Obvious D_3 mylonitic fabrics and folds observed in rocks of the Hollister zone show signs that dextral slip is a possibility along this zone as well.

Bobyarchick (1981) proposed a model in which the Raleigh belt and the similar Kiokee belt of South Carolina and Georgia were tranported along strike-slip regimes in response to oblique convergence of North America and Africa in an Alleghanian collisional event. If, as this study suggests, both the D_2 and D_3 events are Alleghanian, the oblique convergence model provides a viable explanation for a progressive change from east-west compression to dextral movement along north-south trending fault zones.

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REFERENCES CITED

Billings, M.P., and White, W.S., 1950, Metamorphosed mafic dikes of the Woodsville quadrangle, Vermont and New Hampshire: American Mineralogist, v. 35, p. 629-643.

Bobyarchick, A.R., 1981, The eastern Piedmont fault system and its relationship to Alleghanian tectonics in the southern Appalachians: Journal of Geology, v. 89, p. 335-347.

Boltin, W.R., 1985, Geology of the Hollister 7 1/2-minute Quadrangle, Warren and Halifax Counties, North Carolina: Metamorphic transition in the Eastern Slate belt: M.S. thesis, North Carolina State University, 87 p.

Farrar, S.S., 1980, Geology of the Raleigh block and the adjacent Piedmont of North Carolina: *in* Costain and Glover (eds.), Evaluation and Targeting of Geothermal Energy Resources in the southeastern United States, Progress Report, Oct. 1, 1979-June 30, 1980, p. A-1-A-53.

Farrar, S.S., 1984, The Goochland granulite terrane: remobilized Grenville basement in the eastern Virginia Piedmont: Geological Society of America Special Paper 194, p. 215-227.

Farrar, S.S., 1985a, Tectonic evolution of the easternmost Piedmont, North Carolina: Geological Society of America Bulletin, v. 96, p. 362-380.

Farrar, S.S., 1985b, Stratigraphy of the northeastern North Carolina Piedmont: Southeastern Geology, v. 25, p. 159-183.

Farrar, S.S., Russell, G.S., Russell, C.W., and Glover, L. III, 1981, Alleghanian deformation and metamorphism in the eastern Piedmont of North Carolina: New evidence from Rb-Sr whole-rock and biotite ages: Geological Society of America Abstracts with Programs, v. 13, p. 449-450.

Fullagar, P.D., and Butler, J.R., 1979, 325 to 265 m.y.-old granitic plutons in the Piedmont of the southern Appalachians: American Journal of Science, v. 271, p. 161-185.

Gibson, G.G., 1984, Pteridinium: An element of late Precambrian Ediacaran fauna from the Carolina Slate Belt, southern Appalachian orogen (abs.): American Association of Petroleum Geologists Bulletin, v. 68, p. 479-480.

Harvey, B.W., 1974, The microscopic petrography and ore microscopy of the Boy Scout-Jones molybdenum prospect, Halifax County, North Carolina:

M.S. thesis, North Carolina State University, 85 p.
Hatcher, R.D., Jr., Howell, D.E., and Talwani, P., 1977, Eastern Piedmont fault system: Speculation on its extent: Geology, v. 5, p. 636-640.
King, P.B., 1955, A geologic section across the southern Appalachians: An outline of the geology in the segment in Tennessee, North Carolina, and South Carolina: *in* Russell, R.J. (ed.), Guides to Southeastern Geology: Boulder, Co., Geological Society of America, p. 332-373.
Kite, L.E., and Stoddard, E.F., 1984, The Halifax County complex: Oceanic lithosphere in the eastern North Carolina Piedmont: Geological Society of America Bulletin, v. 95, p. 422-432.
Maher, H.D., Palmer, A.R., Secor, D.T., and Snoke, A.W., 1981, New trilobite locality in the Piedmont of South Carolina and its regional implications: Geology v. 9, p. 34-36.
McDaniel, R.D., 1980, Geologic map of Region K, North Carolina: North Carolina Department of Natural Resources and Community Development, Geological Survey Section, open file map.
Parker, J.M. III, 1968, Structure of easternmost North Carolina Piedmont: Southeastern Geology, v. 9, p. 117-131.
Parker, J.M. III, 1977, Structure of Raleigh belt of eastern Piedmont in Wake County, North Carolina: Geological Society of America Abstracts with Programs, v. 9, p. 173.
Parker, J.M. III, 1978, Structure of west flank of the Raleigh belt, North Carolina: *in* Snoke, A.W. (ed.), Carolina Geological Society Guidebook for 1978 Annual Meeting, Columbia, South Carolina, p. 1-7.
Parker, J.M. III, 1979, Geology and mineral resources of Wake County: North Carolina Department of Natural Resources and Community Development, Geological Survey Section, Bulletin 86, 122 p.
Pearce, J.A., and Cann, J.R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analyses: Earth and Planetary Science Letters, v. 19, p. 290-300.
Pettijohn, F.J., 1975, Sedimentary Rocks: 3d ed., New York, Harper & Row, 628 p.
Russell, G.S., Russell, C.W., and Farrar, S.S., 1985, Alleghanian deformation and metamorphism in the eastern North Carolina Piedmont: Geological Society of America Bulletin, v. 96, p. 381-387.
Secor, D.T., Jr., Samson, S.L., Snoke, A.W., and Palmer, A.R., 1983, Confirmation of the Carolina Slate belt as an exotic terrane: Science, v. 221, p. 649-651.
Sigurdsson, H., Sparks, R.S.J., Carey, S.N., and Huang, T.C., 1980, Volcanogenic sedimentation in the Lesser Antilles Arc: Journal of Geology, v. 88, p. 523-540.
Simpson, C., 1986, Determination of movement sense in mylonites: Journal of Geological Education, v. 34, p. 246-261.
Simpson, C., and Schmid, S.M., 1983, An evaluation of criteria to deduce the sense of movement in sheared rocks: Geological Society of America Bulletin, v. 94, p. 1281-1288.
Spry, A., 1969, Metamorphic Textures: Oxford, Pergamon Press, 350 p.
St. Jean, J., Jr., 1973, A new Cambrian trilobite from the Piedmont of North Carolina: American Journal of Science, v. 273-A, p. 196-216.
Stoddard, E.F., and McDaniel, R.D., 1979, Geology of the Raleigh belt in eastern Franklin and Warren Counties, North Carolina: A preliminary report: Geological Society of America Abstracts with Programs, v. 11, p. 214.
Stoddard, E.F., Kite, L.E., and Moye, R.J., 1982, Ophiolitic terranes of the eastern Piedmont of North Carolina and their tectonic implications: Geological Society of America Abstracts with Programs, v. 14, p. 86.
Walker, R.G., 1984, Turbidites and associated coarse clastic deposits, in Walker, R.G. (ed.), Facies Models, Second Edition: Geoscience Canada, Reprint Series 1, Geological Association of Canada, p. 171-188.
Williams, H., 1978, Tectonic lithofacies map of the Appalachian Orogen:

Memorial University of Newfoundland Map No. 1, scale 1:1,000,000.
Williams, H., Turner, F.J., and Gilbert, C.M., 1982, Petrography: An Introduction to the Study of Rocks in Thin Section: San Francisco, W.H. Freeman and Co., 2d ed., 626 p.
Wilson, W.F., 1981, Geologic map of the Halifax and Nash County portion of Region L: North Carolina Department of Natural Resources and Community Development, Geological Survey Section, open file map.
Wylie, A.S., Jr., 1984, Structural and metamorphic geology of the Falls Lake area, Wake County, North Carolina: M.S. thesis, North Carolina State University, 79 p.
Wylie, A.S., Jr., and Stoddard, E.F., 1984, The Falls Lake mylonite zone: A pre-Alleghanian lithotectonic boundary in the eastern North Carolina Piedmont: Geological Society of America Abstracts with Programs, v. 16, p. 702.

APPENDIX: SAMPLE LOCATIONS

From Figure 2

1. Felsic orthogneiss from Bobbitts Branch, approximately one km west of S.R. 1510, northeastern corner of Inez quadrangle.
2. Pelitic schist in unnamed north-flowing tributary to Reedy Creek, west of a point on S.R. 1512, 0.5 km north of its intersection with S.R. 1515 (Hollister quadrangle).
3. Layered metasiltstone-metagraywacke unit, discontinuous roadcut on south side of S.R. 1515, 0.8 km east of its intersection with S.R. 1512.
4. Amphibolite and greenstone, roadcut on S.R. 1521, just south of Reedy Creek, central Hollister quadrangle.
5. Metamudstone-metagraywacke in loose rocks in abandoned quarry located on an unnamed dirt road that extends northward from S.R. 1322 from a point approximately one km south of Little Fishing Creek, southeastern Hollister quadrangle.
6. Graded volcanic metaconglomerate in roadside ditch on north side and just east of intersection of S.R. 1327 and S.R. 1002 near downtown Hollister.
7. Felsic metavolcanic in north-flowing tributary to Little Fishing Creek, approximately 0.3 km south of S.R. 1002, west of its intersection with S.R. 1328, southeastern Hollister quadrangle.
8. Amygdaloidal metabasalt, in same creek as sample 7, about 0.5 km north of S.R. 1002.
9. Metagabbro, in Little Fishing Creek, 0.4 km southeast of the intersection of S.R. 1315 with N.C. 561.
10. Metadiorite in Little Fishing Creek, 0.5 km west of S.R. 1322.
11. Weakly graded metasiltstone-metagraywacke, in Gunter's Creek, 0.25 km north of S.R. 1634, eastern Inez quadrangle.
12. Metamudstone in roadside ditch, east edge of S.R. 1322, 50 m south of Little Fishing Creek, southeastern Hollister quadrangle.

From Table 1

1. HRB-260, metabasalt from outcrop 30 m downstream from amygdaloidal metabasalt, sample 8 above.
2. HES-K2, amphibolite from south-flowing tributary 0.2 km north of Little Fishing Creek and 0.4 km west of S.R. 1509, near northern edge of Hollister quadrangle ("T2" on Figure 2).
3. HRB-237: same as sample 9 above.
4. HRB-256: same as sample 10 above.

[illegible]

[illegible] SAMPLE LOCATIONS

From Figure 1

1. [illegible] corner of [illegible]
2. [illegible]
3. [illegible]
4. [illegible]
5. [illegible] from S.R. [illegible]
6. [illegible] north side [illegible]
7. [illegible] intersection with [illegible]
8. [illegible]
9. [illegible]
10. [illegible]
11. [illegible]
12. [illegible]

From Table 1

1. [illegible]
2. [illegible]
3. [illegible]
4. [illegible]

FOSSIL PLANTS FROM MISSISSIPPIAN-PENNSYLVANIAN TRANSITION STRATA IN THE SOUTHERN APPALACHIANS

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ABSTRACT

Fossil plants, including some not known previously in North America, have been found in the lower-middle part of the Parkwood Formation near Birmingham, Alabama. They indicate an age that is younger than the age of the youngest Mississippian, but older than the age of strata recognized as Pennsylvanian. The plant-bearing beds are equivalent in age to strata placed in the Namurian in Europe. The Parkwood strata in the Birmingham area may form a continuous succession across the Mississippian-Pennsylvanian boundary.

INTRODUCTION

The Parkwood Formation of Alabama is part of a southwestward-thickening, northeastward-prograding foreland clastic wedge (Thomas, 1972, 1974). Regional stratigraphy of the clastic wedge is known from outcrops in the Appalachian fold-thrust belt (Valley and Ridge) and from numerous wells in the Black Warrior basin (Figure 1). The Parkwood consists mainly of deltaic to shallow-marine sandstones and mudstones. The formation overlies Upper Mississippian strata of the Floyd Shale and Bangor Limestone, and underlies the Lower Pennsylvanian Pottsville Formation which consists of barrier-island sandstones and delta-plain sandstones, mudstones, and coals (Figure 2). Both the base and top of the northeastward-prograding Parkwood Formation are regionally diachronous (Thomas, 1972, 1974).

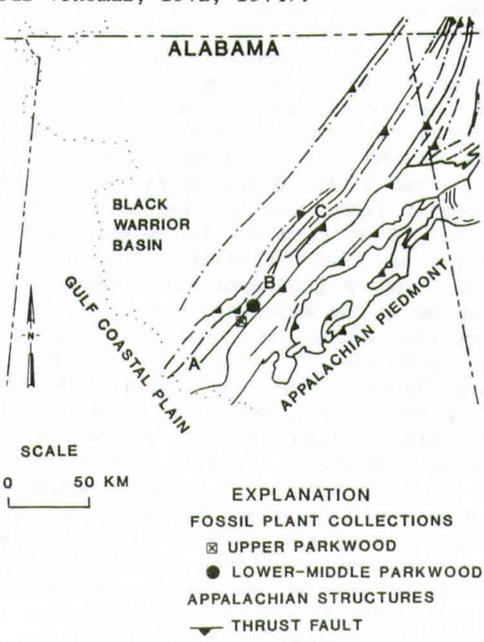

Figure 1. Map of Alabama showing collecting localities in Parkwood Formation in Cahaba synclinorium. Line A-B-C shows trace of cross section of Figure 2.

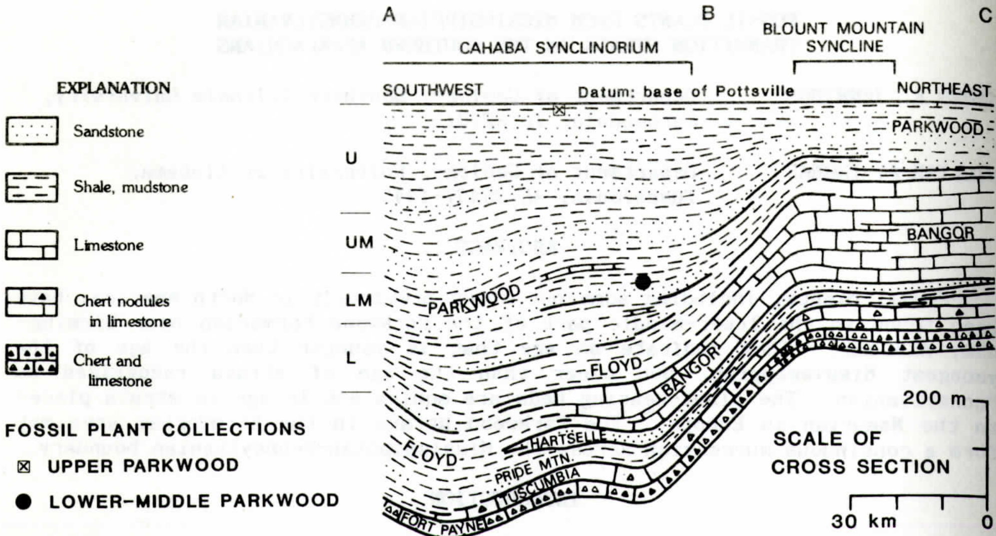

Figure 2. Stratigraphic cross section showing positions of collecting localities in Parkwood Formation in Cahaba synclinorium. Line of cross section shown on Figure 1. The four cycles within the Parkwood clastic sequence are designated lower (L), lower-middle (LM), upper-middle (UM), and upper (U).

The age of the Parkwood Formation is somewhat uncertain, and the unit has been referred to both the Pennsylvanian and the Mississippian (Butts, 1926, 1927; Moore and others, 1944; Culbertson, 1963; Thomas, 1972; Pfefferkorn and Gillespie, 1982). Fossils are generally rare in the Parkwood Formation (Thomas, 1972), and many of the recorded fossils either are not diagnostic or belong to assemblages that have affinities with both systems. Fossil plant lists from the upper part of the Parkwood Formation (Moore and others, 1944; Pfefferkorn and Gillespie, 1982) have suggested a Pennsylvanian age for that part of the unit, but the plants have not been either figured or described. A few fossil plants and invertebrates (listed in an abstract, but not figured or described) from one of the few exposures of the Parkwood in northwestern Alabama along the north limb of the Black Warrior basin have been interpreted to indicate a middle Early Pennsylvanian age (Henry and others, 1981). In regional context, however, that locality is in the stratigraphically highest and most distal fringe of the northeastward-prograding Parkwood clastic sequence (Thomas, 1972), and is geographically remote from Parkwood outcrops in the Cahaba synclinorium in the Appalachian fold-thrust belt (Figure 1).

In a linear outcrop belt along the Cahaba synclinorium, deltaic facies of the Parkwood Formation prograde northeastward over the prodelta Floyd Shale and the shallow-marine Bangor Limestone (Figure 2) (Thomas, 1972). The Parkwood consists of a cyclic succession of sandstones and mudstones that represent delta-plain, distributary, marsh, and marine-bay settings. Thin marine units, including limestone, locally contain invertebrate fossils. Delta-plain sequences include thin coal beds, carbonaceous mudstones, and plant fossil-bearing mudstones. Some mudstones contain well preserved rooted zones preserved *in situ*. The cyclicity of the Parkwood reflects progradation of successive delta lobes, and along the Cahaba synclinorium, four large-scale cycles are reflected in the succession of sandstones and mudstones (Figure 2) (Thomas, 1974). The four Parkwood cycles are designated as lower, lower-middle, upper-middle, and upper.

COLLECTING LOCALITY

Construction of Interstate 20 near Birmingham exposed a mudstone containing abundant fossil plants in the lower-middle part (second cycle above the base) of the Parkwood Formation in the Cahaba synclinorium. The locality is in the NW 1/4, NE 1/4, Sec. 25, T. 17 S., R. 2 W., Jefferson County, Alabama (Irondale 7.5-minute Quadrangle) (Figure 1). The fossil-bearing bed is approximately 185 m (355 ft) above the base and 364 m (1195 ft) below the top of the Parkwood (Figure 2). A large collection was made by the authors during construction of Interstate 20 in 1981, and in addition, a reference collection was deposited at the Geological Survey of Alabama. The specific collecting site was destroyed during completion of Interstate 20, but the fossil-bearing beds are preserved in the southern corner of the interchange at Exit 133.

BIOSTRATIGRAPHY

Preliminary investigation of the fossil plants from this site shows many stratigraphically diagnostic forms (Figures 3-8) that are younger than any strata of the type Mississippian, but are older than strata of the Pocahontas Formation of West Virginia (Table 1). The Pocahontas is generally regarded as the earliest Pennsylvanian. Three of the fossil plants identified from the lower-middle Parkwood Formation are forms that also occur near the top of the Mississippian in the type region. Seven of the forms recognized in the lower-middle Parkwood Formation also occur in strata that are generally recognized as Pennsylvanian in the eastern United States. Eight of the fossil plant forms from the lower-middle Parkwood Formation are not known from either Mississippian or Pennsylvanian strata of the eastern United States. *Pecopteris* "*aspera*" has been reported from beds in West Virginia and Virginia that are assigned to the Mississippian (Pfefferkorn and Gillespie, 1982). However, investigation of sterile foliage of *Senftenbergia* in the type area of the Mississippian in Illinois (Jennings and Eggert, 1977) has shown that some Mississippian forms are very similar to *P. aspera*, but differ from the

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Table 1. List of stratigraphically significant fossil plants from the lower-middle part of the Parkwood Formation near Birmingham, Alabama, in comparison with their occurrence in known Mississippian and Pennsylvanian strata elsewhere in North America. Fossil plants not listed as occurring in either the type Mississippian or the Pennsylvanian have not been reported previously from North America.

Parkwood Formation along Interstate 20 near Birmingham, Alabama	Type Mississippian	Pennsylvanian (Pennsylvania, Illinois, West Virginia, Virginia, Kentucky, Colorado)
Lepidodendron aculeatum	no	yes
Sigillaria elegans	no	yes
Calamites baldurnensis	no	no
C. cistiiformis	yes	yes
Asterophyllites longifolius	no	yes
Calamostachys ramosa	no	no
Sphenophyllum tenerrimum	yes	no
Bowmanites sp.	yes	yes
Alloiopteris dumontii	no	no
Pecopteris (Senftenbergia) aspera	no	no
Sphenopteris baldurnense	no	no
S. schatzlarensis	no	yes
S. stochesianum	no	no
Dicranophyllum sp.	no	yes
Calathiops beinertiana S. & W. (*non* Goeppert)	no	no

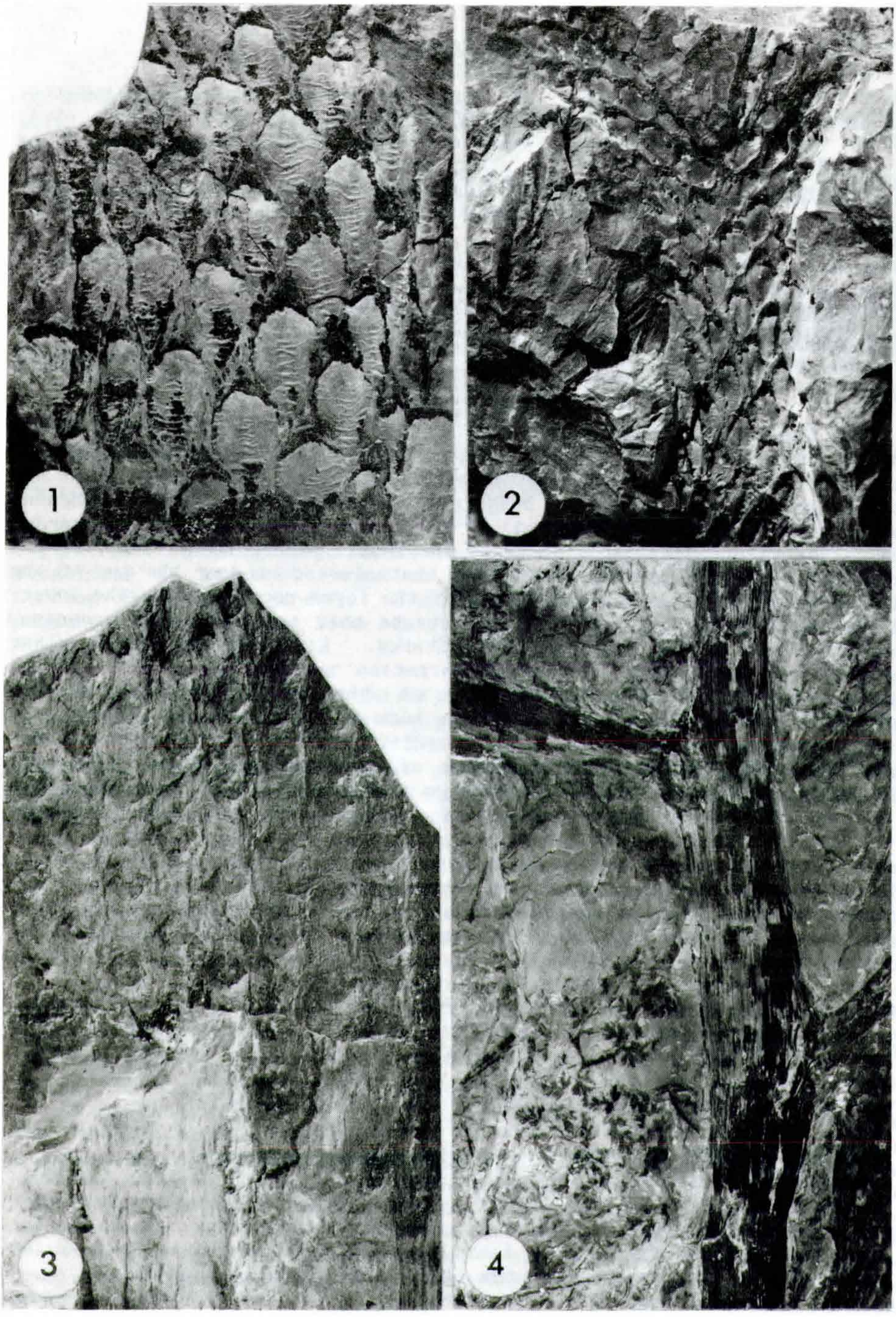

Figure 3. Fossil plants from the Parkwood Formation.

1, *Lepidodendron aculeatum*, stem surface. X1.1.

2, *Lepidodendron aculeatum*, a small bifurcating branch. X1.1.

3, *Calamites cistiiformis*. X0.55.

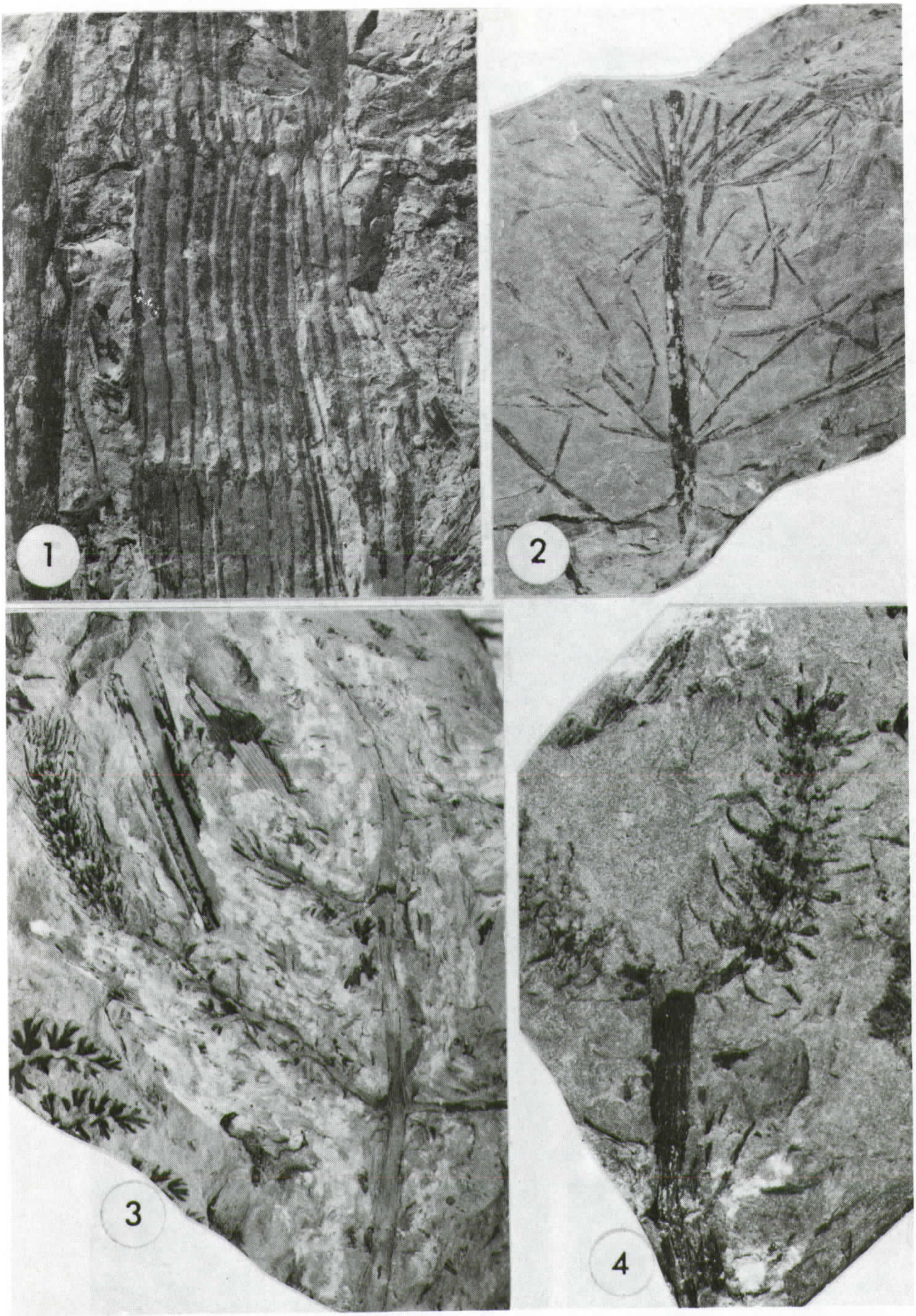

Figure 4. Fossil plants from the Parkwood Formation.

1, *Calamites baldurnense*. X1.1.

2, *Asterophyllites longifolius*. X1.1.

3, Specimen bearing *Calamostachys*-type cone and *Asterophyllites*-type foliage. X1.1.

4, *Calamostachys ramosa*. X2.

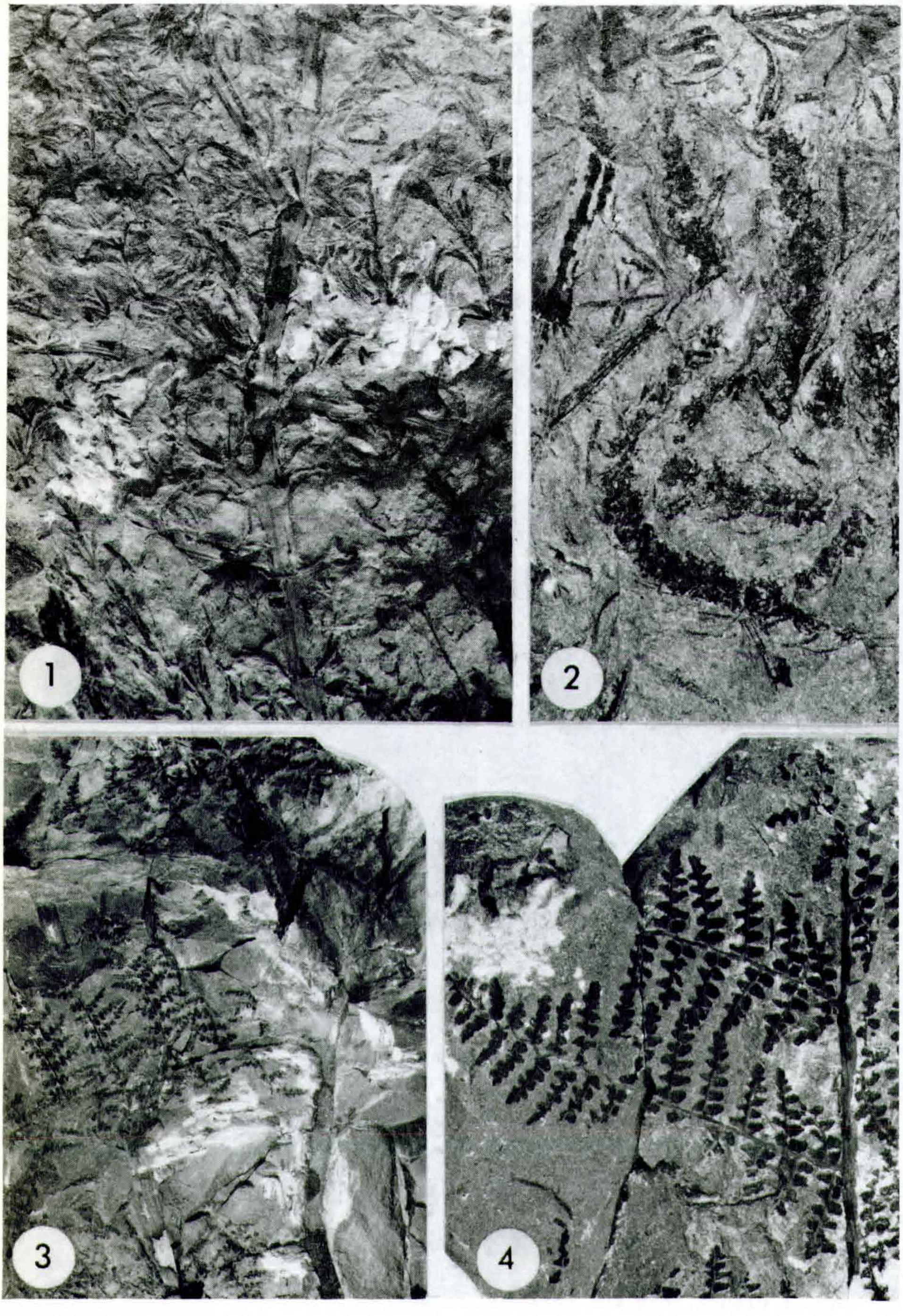

Figure 5. Fossil plants from the Parkwood Formation.

1, *Sphenophyllum tenerrimum*. X1.1.

2, *Bowmanites* sp. X2.2.

3, *Pecopteris aspera*. X0.55.

4, *Pecopteris aspera*. X1.1.

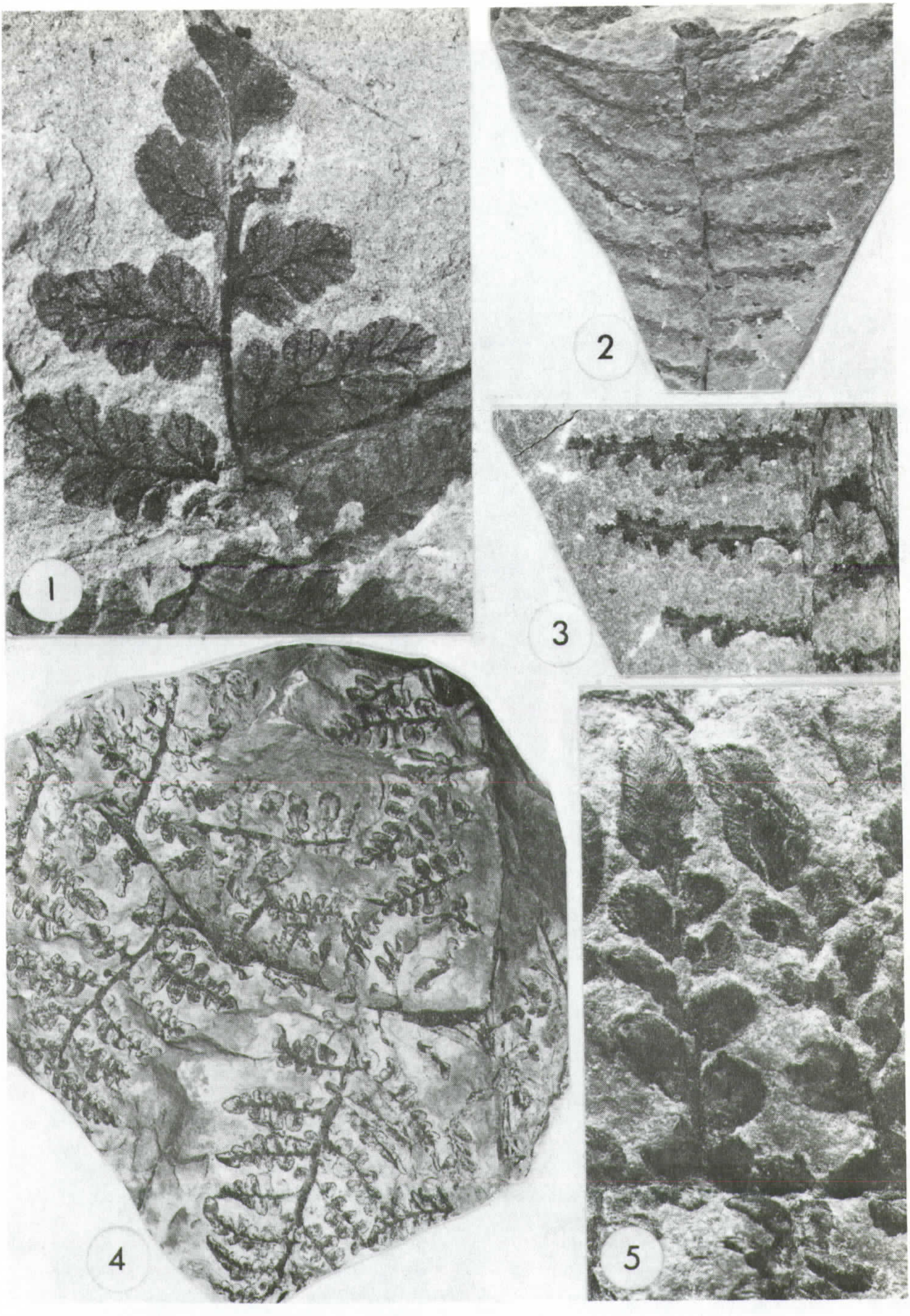

Figure 6. Fossil plants from the Parkwood Formation.
1, *Pecopteris aspera*. X4.4.
2, *Alloiopteris dumontii*. X1.1.
3, *Alloiopteris dumontii*. X2.2.
4, *Neuropteris bulupalganensis*. X1.1.
5, *Neuropteris bulupalganensis*. X4.4.

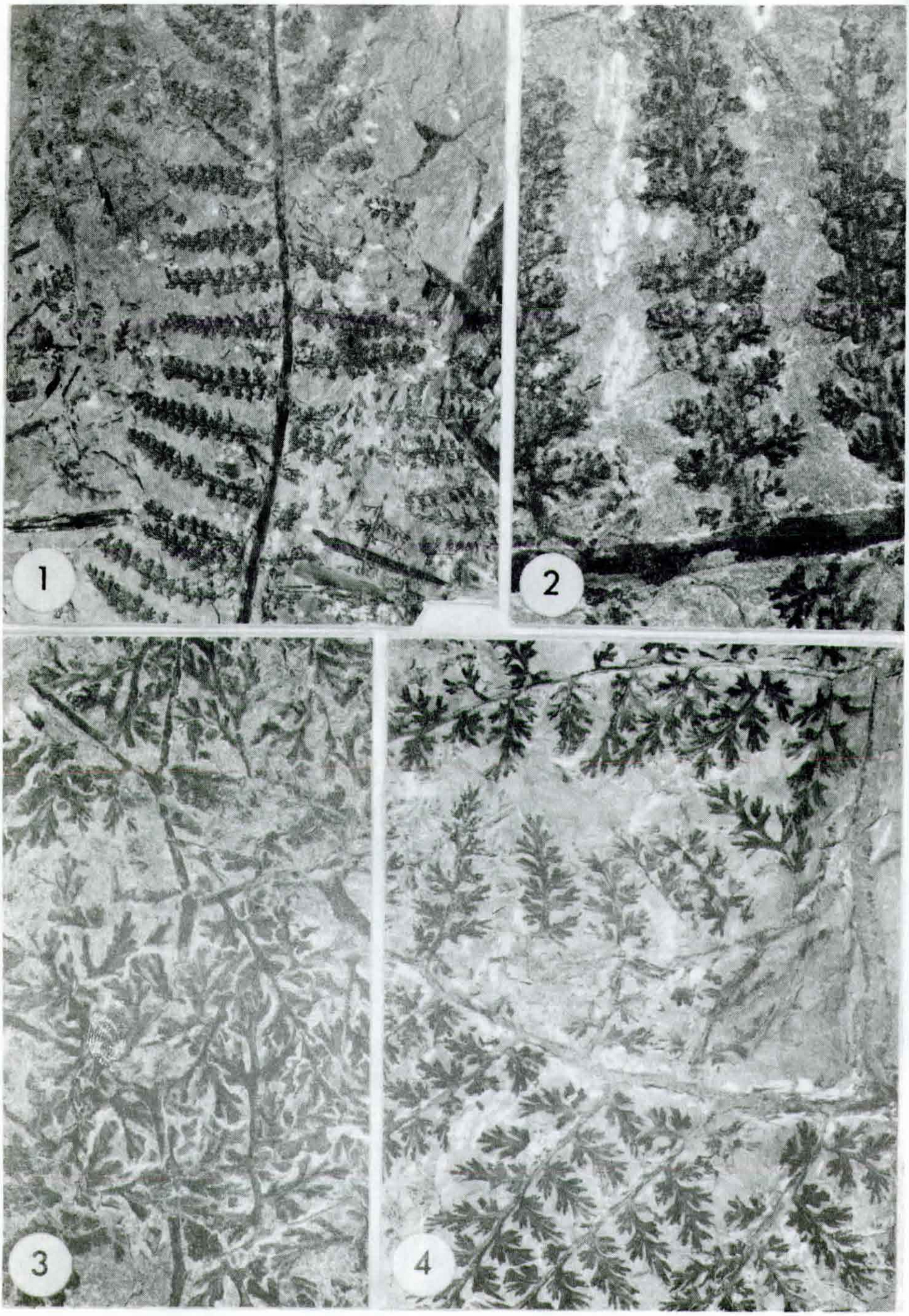

Figure 7. Fossil plants from the Parkwood Formation.
1, *Sphenopteris baldurnense*. X0.55.
2, *Sphenopteris baldurnense*. X2.2.
3, *Sphenopteris schatzlarensis*. X1.1.
4, *Sphenopteris stochesianum*. X1.1.

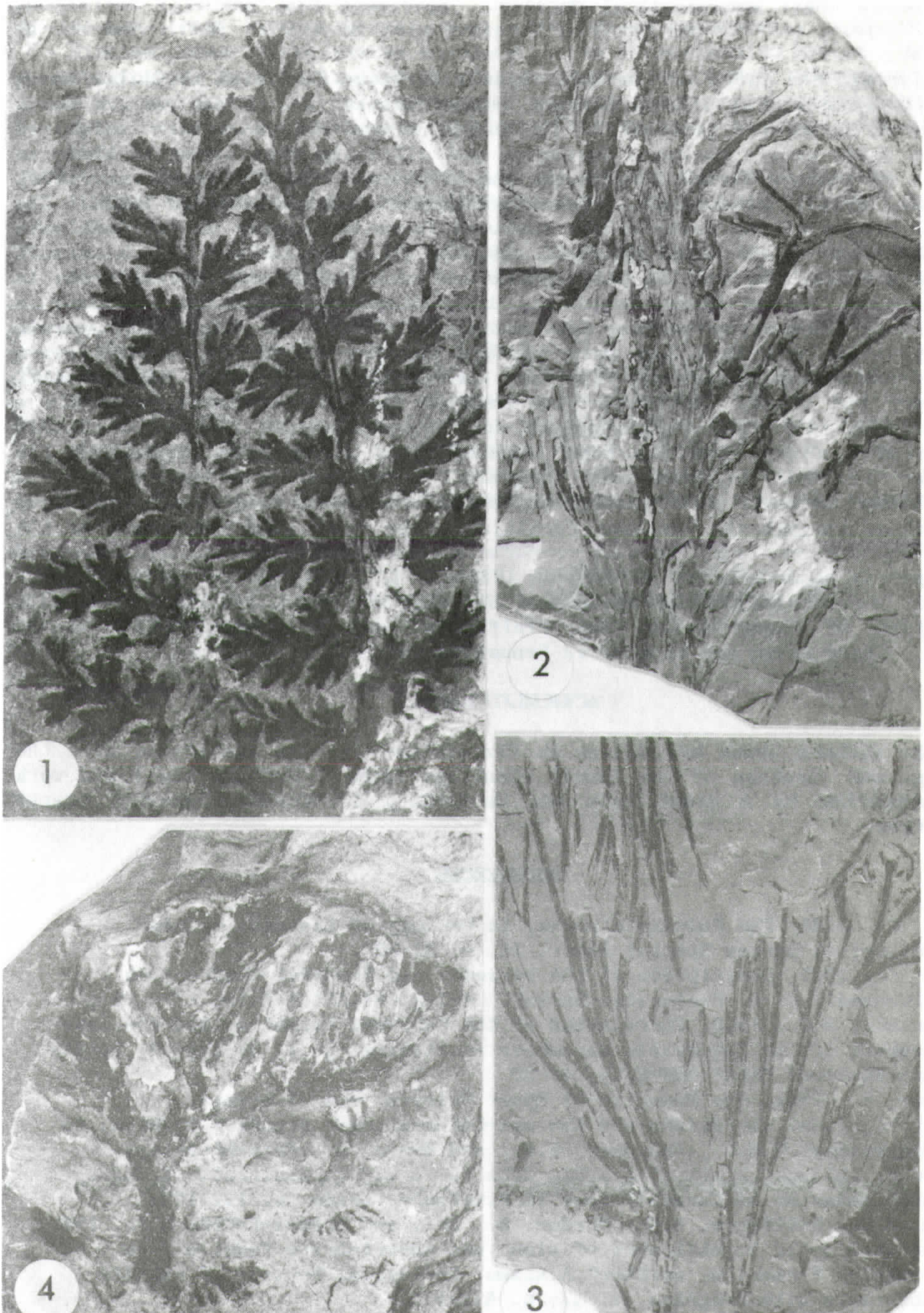

Figure 8. Fossil plants from the Parkwood Formation.
1, *Sphenopteris stochesianum*. X2.2.
2, *Dicranophyllum* sp. X1.1.
3, *cf. Dicranophyllum sp*. X1.1.
4, *Calathiops beinertiana* Stockmans (*non* Goeppert). X2.2.

species in small morphological details. Specimens that actually correspond to *P. aspera* are not known from Mississippian strata of North America at present.

Although the flora of the lower-middle Parkwood Formation is unique among Carboniferous floras of the eastern United States, it is not unique among floras of Euramerica generally. The fossil plants that have been recognized in the lower-middle Parkwood Formation are all forms that occur in Namurian strata of western Europe (Stockmans and Williere, 1952-53; Purkynova, 1970; Havlena, 1982), and include several particularly characteristic taxa. The fossil plants of the lower-middle Parkwood Formation appear to fit best with a correlation to the lower Namurian, possibly E_1 or E_2.

At present, the section in Virginia and West Virginia is the only one in the eastern United States that has been widely believed to represent a continuous succession across the boundary between the Mississippian and Pennsylvanian and that has been studied paleobotanically in recent years (Pfefferkorn and Gillespie, 1982). As Wagner (1982) has pointed out, however, the fossil plants in that section indicate a hiatus, probably a substantial one. The fossil plants of the lower-middle Parkwood Formation represent a flora from the time interval that is not represented in the section in West Virginia and Virginia. Physical stratigraphy of the prograding deltaic and transgressive marine units suggests continuous deposition through the Parkwood Formation. The upper part (uppermost of the four cycles) of the Parkwood contains *Sphenopteris hoeninghausii* and *Neuropteris pocahontas* at Birmingham (Figures 1, 2) and cannot be older than floras in the Pocahontas Formation of West Virginia and the Pottsville Formation of Pennsylvania (see Pfefferkorn and Gillespie, 1982; White, 1900). These units are considered Pennsylvanian.

ACKNOWLEDGMENTS

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REFERENCES

Butts, C., 1926, The Paleozoic rocks, *in* Geology of Alabama: Alabama Geological Survey Special Report 14, p. 41-230.

Butts, C., 1927, Description of the Bessemer and Vandiver Quadrangles (Alabama): U. S. Geological Survey, Geological Atlas Folio 221, 22 p.

Culbertson, W.C., 1963, Correlation of the Parkwood Formation and the lower members of the Pottsville Formation in Alabama: U. S. Geological Survey Professional Paper 450-E, p. 47-50.

Havlena, V., 1982, The Namurian of Upper Silesia: Floral breaks, lithological variations and the mid-Carboniferous boundary problem, *in* Ramsbottom, W.H.C., Saunders, W.B., and Owens, B., Paleobotanical contributions; Biostratigraphic data for a mid-Carboniferous boundary: Subcommission on Carboniferous Stratigraphy, Leeds, p. 112-119.

Henry, T.W., Gillespie, W.H., Gordon, M., Jr., and Schweinfurth, S.P., 1981, Stratigraphic significance of plant and invertebrate fossils from Parkwood Formation, northern Alabama: Geological Society of America Abstracts with Programs, v. 13, p. 471-472.

Jennings, J.R., and Eggert, D.A., 1977, A preliminary report on permineralized *Senftenbergia* from the Chester Series of Illinois: Review of Paleobotany and Palynology, v. 24, p. 221-225.

Moore, R.C., and others, 1944, Correlation of Pennsylvanian formations of North America: Geological Society of America Bulletin, v. 55, p. 657-706.

Pfefferkorn, H.W., and Gillespie, W.H., 1982, Plant megafossils near the Mississippian-Pennsylvanian boundary in the Pennsylvanian System stratotype, West Virginia/Virginia, *in* Ramsbottom, W.H.C., Saunders, W.B., and Owens, B., Palaeobotanical contributions; Biostratigraphic data for a mid-Carboniferous boundary: Subcommission on Carboniferous Stratigraphy, Leeds, p. 128-133.

Purkynova, E., 1970, Die Unternamurflora des Beckens von Horni Slezcko (CSSR): Palaontologische Abhandlungen, Abteilung B, Band III, Heft 2, p. 129-268.

Stockmans, F., and Williere, Y., 1952-53, Vegetaux Namuriens de la Belgique: Association pour l'Etude, de la Paleontologie et de la Stratigraphie Houilleres Publication 13, 382 p.

Thomas, W.A., 1972, Mississippian stratigraphy of Alabama: Alabama Geological Survey Monograph 12, 121 p.

Thomas, W.A., 1974, Converging clastic wedges in the Mississippian of Alabama: Geological Society of America Special Paper 148, p. 187-207.

Wagner, R.H., 1982, Floral changes near the Mississippian-Pennsylvanian boundary; an appraisal, *in* Ramsbottom, W.H.C., Saunders, W.B., and Owens, B., Palaeobotanical contributions; Biostratigraphic data for a mid-Carboniferous boundary: Subcommission on Carboniferous Stratigraphy, Leeds, p. 120-127.

White, D., 1900, The stratigraphic succession of the fossil floras of the Pottsville Formation in the Southern Anthracite Coal Field, Pennsylvania: U. S. Geological Survey 20th Annual Report, pt. 2, p. 749-930.

RECOGNITION OF A PROTEROZOIC CAULDRON BOUNDARY IN SOUTHEASTERN MISSOURI

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ABSTRACT

The St. Francois Mountains, Missouri, are a southwesterly tilted (10^o-15^o) exposure of an extensive mid-continent basement accretionary terrane of 1440-1480 Ma epizonal granites and rhyolites. A regionally exposed angular unconformity, interpreted to be a cauldron topographic boundary, has been mapped in detail for a distance of 13 km in the central part of the Proterozoic St. Francois Mountains terrane. Along the intracáldera side of the angular unconformity, lenses of megabreccia and mesobreccia are completely enveloped by younger ignimbrites. Several lava domes and an elongate granite porphyry stock are spatially associated with the angular unconformity and intrude buried ring fractures. Although the angular unconformity is in the correct location to warrant correspondence with the proposed Taum Sauk Caldera, an exceedingly thick intracaldera sequence would be required for correlation (>4-6 km).

The cauldron segment represents the only known surface exposure that has been confirmed by detailed field studies in the St. Francois Mountains volcanic section. The abundance of lava domes suggests that the caldera was not extensively eroded prior to burial by younger volcanic rocks. It seems probable that much of the epizonal granite and rhyolite must have resulted from collapse caldera eruptions.

INTRODUCTION

The presence of an exposed cauldron in the Proterozoic St. Francois Mountains of Missouri was first proposed by Anderson and others (1969), who interpreted arcuate pluton exposures as ring fracture intrusions along the western margin of the Taum Sauk Caldera. The eastern boundary of the caldera was postulated by Sides and others (1981). Due to the excessively thick (4-6 km) intracaldera sequence of tuffs and the potential for caldera nesting, the Taum Sauk Caldera name was not used by Sides and others (1981). They also established the presence of an exposed granite ring complex in the predominantly intrusive part of the terrane which they identified as the highly eroded Butler Hill Caldera. Several buried cauldrons have been postulated on the basis of geophysical anomalies within the granite-rhyolite terrane (Cordell, 1979; Kisvarsanyi, 1980, 1981).

Eroded remnants of Proterozoic calderas are also exposed in the Arabian Shield (Dodge, 1979) and near Boston, Massachusetts (Thompson, 1985). Folded Proterozoic cauldrons have been described for the 1.9 Ga Wopmay Orogen (Hildebrand, 1984). Silver and others (1986) recently described a 1700 Ma alkali-rhyolite caldera and subjacent magma system in central Arizona. These occurrences imply that collapse caldera eruptions were significant during the Proterozoic evolutionary history of these regions.

The purpose of this paper is to provide field and petrologic evidence for a well defined Proterozoic cauldron topographic boundary in the central part of the St. Francois Mountains. In this paper, "topographic boundary" is an eroded equivalent to the caldera "topographic wall" defined by Lipman (1976). Similarities between the proposed boundary and Tertiary caldera boundaries of the western United States are demonstrated. The implications for collapse caldera eruptions in the mid-continent are also discussed.

Geologic Setting

The St. Francois Mountains provide a rare exposure of an extensive subsurface terrane dominated by 1440-1480 Ma rhyolites and epizonal granites (Thomas and others, 1984). The study area shown in Figure 1 occurs in the central part of the exposed terrane. Volcanic rocks are predominantly rhyolitic to rhyodacitic ignimbrites and lava flows. Intrusives include granite and quartz monzonite. Volcanic sheets are very similar in appearance to Tertiary tuffs exposed in the western United States. Although metamorphic fabric has not been detected, preservation of rocks has been enhanced by secondary silicification.

Within the study area, a regional angular unconformity has been mapped for about 13 km and is known to extend an additional 3 km south of the study area, where it is covered by Paleozoic rocks. This is expected given the regional tilt of 10^{o}-15^{o} to the west and southwest for the exposed terrane (Bickford and others, 1977). Along this unconformity, steeply dipping older volcanic rocks located to the east and northeast are in contact with younger gently dipping volcanic rocks to the west and southwest (Figure 2). The younger volcanic units generally dip to the west and southwest approximately 10^{o}-20^{o}.

Rock Types and Stratigraphic Relations

The rocks described in this paper are divided into two groups based on their spatial relationships to the angular unconformity (Figure 1). Rocks exposed east and north of this feature are older than those to the west since western rocks contain xenoliths of eastern rocks. Western rocks generally dip to the southwest approximately 10^{o}-20^{o}. These values are similar to the regional attitude of the St. Francois Mountain terrane (Bickford and others, 1977).

Igneous Rocks Exposed East and North of the Angular Unconformity: Precauldron Rocks

Volcanic rocks include the Lake Killarney Formation and Grassy Mountain Ignimbrite which typically contain abundant alkali feldspar and quartz phenocrysts relative to plagioclase (Table I). The lower member of the Lake Killarney Formation is a crystal-poor rhyolitic tuff with locally abundant secondary flow structures. Phenocrysts include perthitic feldspars (probably sanidine originally), quartz, sparse plagioclase and Fe-Ti oxide. The upper part of the unit is a crystal-poor tuff with abundant xenoliths consisting of rhyolitic tuff, rhyolite lava flows, bedded tuff and diabase. These are interpreted to be derived from depth during emplacement of the sheet. Phenocrysts within the crystal-poor upper member include alkali feldspar, quartz,

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Table I. Modal Data for Precauldron Igneous Rocks

	Lake Killarnary Formation Lower Member (n=3)	Upper Member (n=5)	*Grassy Mountain Ignimbrite* (n=6)	*Butler Hill Granite* (n=26)	*Stono Granite*
Ground mass	*80.4 ± 2.0	93.0 ± 5.0	67.1 ± 7.8	-	-
Quartz	1.7 ± 2.1	1.3 ± 1.2	10.0 ± 5.1	37.0 ± 4.0	24.4
Alkali Feldspar	6.3 ± 1.6	3.0 ± 1.9	11.2 ± 2.3	57.2 ± 7.2	61.4
Oligoclase	<0.1	<0.1	<0.1	4.5 ± 6.3	4.6
Biotite	-	-	<0.1	0.7 ± 0.8	-
Amphibole	-	-	<0.1	-	5.7
Fe-Ti Oxides	<0.1	<0.1	0.6 ± 0.5	0.6 ± 0.5	3.2
Sphene	-	-	-	<0.1	-
Lithics	-	2.3 ± 4.5	<0.1	-	-
Pumice	10.5 ± 0.98	<0.1	9.0 ± 9.0	-	-

*Values in area percent. Minimum of 1000 counts per thin section.

trace amounts of plagioclase and Fe-Ti oxides. Although the Lake Killarney Formation is estimated to be 520 m thick in the southern part of the study area, it is known to be significantly thicker to the south (Sides, 1978; Sides and others, 1981).

Overlying the Lake Killarney Formation is the Grassy Mountain Ignimbrite which consists of a thin locally developed crystal-depleted base and a thick (500 m) homogeneous cooling unit. Perthitic alkali feldspar (11%) and quartz (10%) are always recognizable megascopically while plagioclase, Fe-Ti oxides and altered remnants of biotite and hornblende are rare. Flattened pumice lumps have maximum length to thickness ratios of 16:1 in the more densely welded zones and 4:1 in marginal zones. Near the central part of this extensive sheet, pumice fragments are recognized only by the euhedral nature of the phenocrysts. Similar observations have been described for highly welded Tertiary ignimbrites (Ratte and Steven, 1967; Nusbaum, 1984). The Grassy Mountain Ignimbrite is chemically similar to the Butler Hill Granite and is believed to be the outflow unit for the deeply eroded Butler Hill Caldera (Sides and others, 1981; Bickford and others, 1981). Stono Granite (Table I) is intruded by Butler Hill Granite and probably represents a ring intrusion for the Butler Hill Caldera (Sides and others, 1981).

Volcanic Rocks Exposed West of the Angular Unconformity: Intracaldera Tuffs and Lava Domes

Exposures of these rocks include ignimbrites, tuff-breccias, lavas, volcaniclastic tuffs and subvolcanic plugs. These rocks are mineralogically distinct from the older rocks due to the presence of plagioclase as the dominant phenocryst phase. Alkali feldspar, quartz, Fe-Ti oxides, altered hornblende and biotite are also present in most of the units (Table II).

Wolf Mountain Ignimbrite is exposed adjacent to the angular unconformity and is the oldest in a sequence of gently dipping volcanic units that continues westward for several kilometers. This crystal-rich tuff contains abundant fragmental plagioclase crystals with embayed quartz, perthitic alkali feldspars, Fe-Ti oxides, biotite and hornblende altered to chlorite. The tuff also contains rare cigar-shaped lenticules similar to those described for the Gribbles Creek Tuff, Colorado (Chapin and Lowell, 1979). This tuff is megascopically indistinguishable from the Cedar Bluff Rhyolite described by Berry (1976) and may be part of a dominant volume system similar to those described by Smith (1979).

Rather poorly defined contact relations suggest that the Iron Mountain Lake Ignimbrite overlies the Wolf Mountain Ignimbrite. This crystal-poor tuff contains abundant flattened pumice fragments which are green due to selective propyllitic alteration. The matrix is black to dark gray due to partial alteration to magnetite and goethite. Some outcrops are sufficiently magnetite-rich that compass needle deflection may be observed. Xenoliths of uncorrelated clasts of diabase, rhyolite and graphic granite are locally abundant. This unit has a minimum thickness of 200 m.

Tribby Breccia conformably overlies Iron Mountain Lake Ignimbrite. This crystal-poor ignimbrite contains abundant recrystallized volcanic xenoliths, probably derived from depth, and a matrix that has been extensively recrystallized to silica and feldspar. Rare pumice fragments exhibit compaction ratios of about 4:1. Maximum thickness of the tuff is 210 m.

A series of locally autobrecciated lava flows with minor volcaniclastic tuff and lahar, known as the Ironton Hollow Rhyolite, overlies Tribby Breccia in the western part of the study area. Ironton Hollow Rhyolite lavas are crystal-poor and exhibit flow banded fabric which is crudely parallel to attitudes displayed by surrounding ignimbrites. Intercalated lahars and volcaniclastic tuffs are thinly-bedded to massive and contain xenoliths which are extensively recrystallized due to nearby iron mineralization. Plagioclase euhedra with minor quartz and Fe-Ti oxides are present in the lavas. In the north central part of the study area, an elongated granite

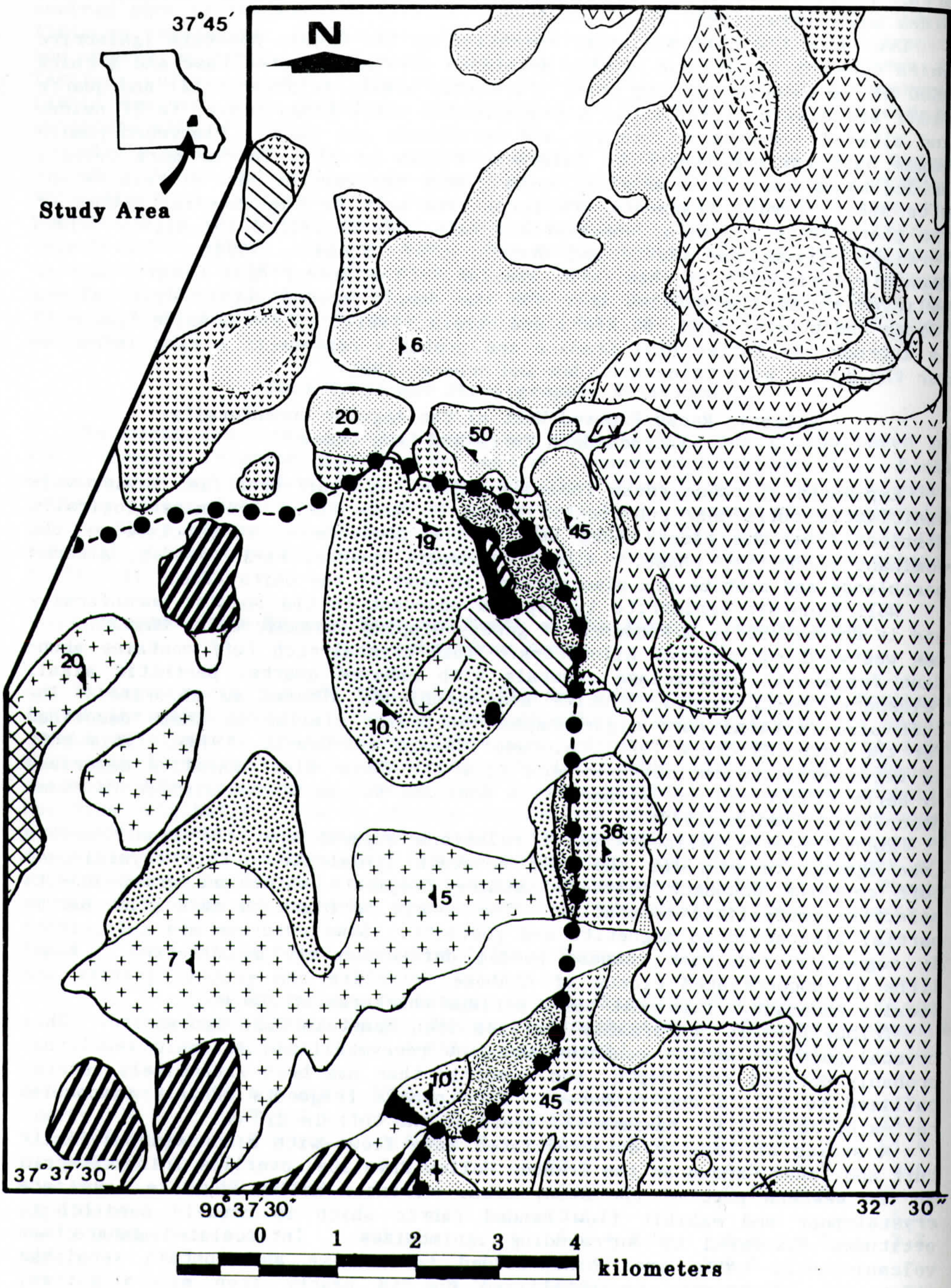

Figure 1. Geology of parts of the Iron Mountain Lake and Graniteville quadrangles, central St. Francois Mountains.

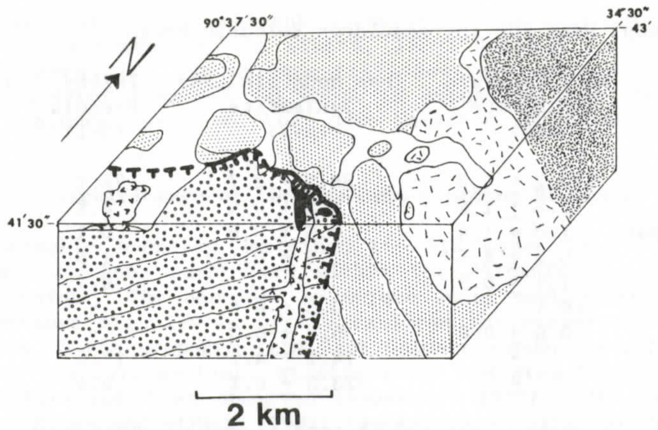

Figure 2. Block diagram showing angular unconformity (hachures) between steeply dipping older rocks (dot pattern; black where they form inclusions in younger rocks) and gently dipping younger rocks (small circles); ring dike and lava dome (v pattern); Butler Hill Granite (large dashes); Stono Granite (small dashes); Paleozoic sedimentary rocks and Quaternary alluvium unpatterned.

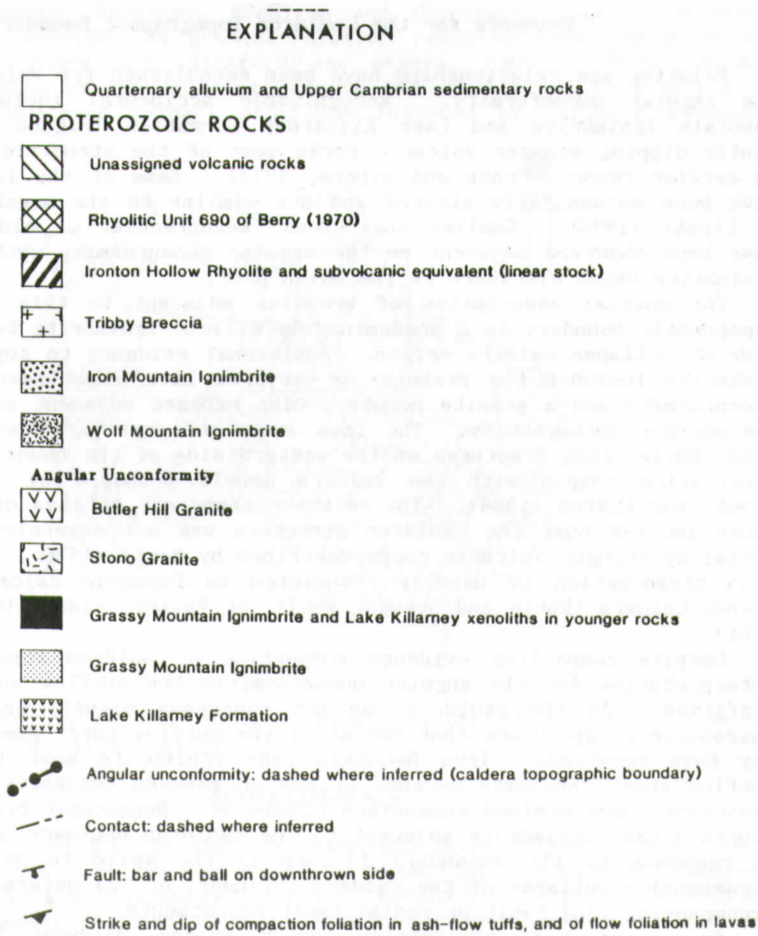

Table II. Modal Data for Intracauldron Volcanic Rocks

	Wolf Mountain Ignimbrite (n=4)	*Iron Mountain Lake Ignimbrite* (n=2)	*Tribby Breccia* (n=1)	*Ironton Hollow Rhyolite* (n=3)
Groundmass	*80.2 ± 4.3	58.9 ± 2.7	74.1	96.7 ± 2.9
Quartz	1.5 ± 1.6	0.5 ± 0.4	0.1	0.9 ± 0.5
Alkali Feldspar	1.3 ± 1.4	0.6 ± 0.8	0.1	<0.1
Oligoclase	15.2 ± 4.5	1.9 ± 0.1	1.9	1.0 ± 0.5
Biotite	1.1 ± 0.9	-	0.7	<0.1
Amphibole	<0.1	0.5 ± 0.5	-	-
Fe-Ti Oxides	0.6 ± 0.6	0.4 ± 0.3	-	<0.1
Lithics	-	14.1 ± 4.1	17.6	<0.1
Pumice	-	23.3 ± 0.1	5.4	-

*Values in area percent. Minimum of 1000 counts per thin section. Ironton Hollow Rhyolite data from Sides, 1978.

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porphyry stock containing quartz, myrmekite, and abundant spherulites is interpreted to be a shallow subvolcanic equivalent to the Ironton Hollow Rhyolite lavas.

Evidence for the Cauldron Topographic Boundary

Relative age relationships have been established for volcanic rocks along the angular unconformity. Recognizable accidental inclusions of Grassy Mountain Ignimbrite and Lake Killarney Formation (Figure 1) occur within gently dipping younger volcanic rocks west of the structure, in contrast to an earlier report (Pratt and others, 1979). Some of the largest inclusions have been mechanically altered and are similar to the megabreccia described by Lipman (1976). Smaller inclusions (mesobreccia) of older volcanic rock have been observed adjacent to the angular unconformity within Wolf Mountain Ignimbrite which elsewhere is inclusion-poor.

The spatial association of breccias adjacent to this regional arcuate topographic boundary in a predominantly silicic ignimbrite terrane is suggestive of collapse caldera origin. Additional evidence to support this interpretation includes the presence of several lava domes next to the angular unconformity and a granite porphyry dike exposed adjacent and subparallel to the angular unconformity. The lava domes and granite porphyry dike may intrude buried ring fractures on the western side of the boundary. These characteristics comply with the caldera models proposed by Smith and Bailey (1968) and Elston (1984). The relative abundance of lava domes to intrusive rocks implies that the cauldron structure was not severely eroded prior to burial by younger volcanic rocks described by Berry (1976). High-level cauldron preservation is usually restricted to Cenozoic calderas such as the Creede Caldera (Ratte and Steven, 1967), or Valles Caldera (Smith and others, 1961).

Despite compelling evidence supporting a cauldron topographic boundary interpretation for the angular unconformity, the outflow sheet has not been confirmed. As the cauldron has not undergone signficiant erosion, it is reasonable to speculate that burial of the outflow tuff beneath younger tuffs may have occurred. Iron Mountain Lake Ignimbrite must be considered an outflow sheet candidate because it has an exposed thickness of 200 m and an additional undetermined subsurface thickness. Megascopic brecciation of Iron Mountain Lake Ignimbrite adjacent to the cauldron boundary must have occurred in response to (1) secondary flowage in the soild to nearly solid state accompanying collapse of the caldera cylinder, or (2) deformation of the tuff accompanying ring fault or radial fault adjustments.

As suggested in preliminary investigations (Nusbaum, 1981), the topographic boundary is in the correct location to correspond to the proposed Taum Sauk Caldera (Anderson and others, 1969) located to the west of the study area (Figure 3). The caldera fill, however, would have an estimated thickness of greater than 4-6 km which is excessive even when compared to much larger calderas such as the Toba Cauldron of Sumatra (Chesner and others, 1983).

Significance of the Explosed Cauldron Boundary

The topographic boundary described in this paper is the only cauldron boundary that has been identified within the volcanic section of the St. Francois Mountains. The boundary is unique among most Proterozoic cauldron boundaries because high levels of the cauldron are preserved. Although the western margin of the proposed Taum Sauk Caldera (Anderson and others, 1969) occurs in the volcanic section, stratigraphic and structural changes across the boundary have not been observed (Anderson, 1970). The proposed Butler Hill Caldera (Sides and others, 1981) represents a highly eroded caldera, or granite ring complex; thus, cauldron collapse evidence has been removed by erosion.

The implications for collapse caldera eruptions in the evolution of the St. Francois Mountains rhyolitic and granitic terrane are significant. The St. Francois Mountains, however, represent a rare exposure of a predominantly subsurface terrane extending beneath Illinois, Indiana, and parts of Missouri, Tennessee, Kentucky, Michigan and Wisconsin (Thomas and others, 1984). This 1440-1480 Ma granite-rhyolite resulted from remelting of a juvenile accretionary arc (Bickford and others, 1986). Thus, it seems probable that much of the epizonal granite and rhyolite of this terrane has resulted from collapse caldera eruptions followed by variable amounts of erosion.

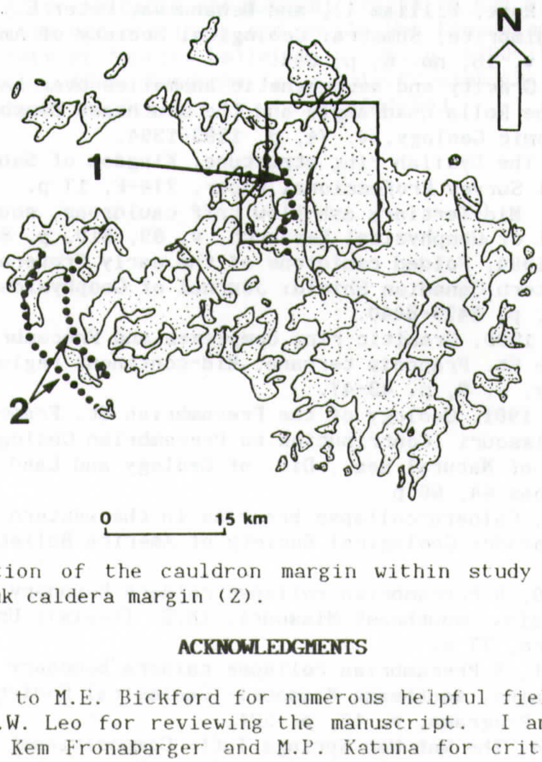

Figure 3. Location of the cauldron margin within study area (1) and the proposed Taum Sauk caldera margin (2).

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REFERENCES CITED

Anderson, J.E., Bickford, M.E., Odom, A.L., and Berry, A.W., 1969, Some age relations and structural features of the Precambrian volcanic terrane, St. Francois Mountains, southeastern Missouri: Geological Society of America Bulletin, v. 80, p. 1815-1818.

Anderson, R.E., 1970, Ash-flow tuffs of Precambrian age in southeast Missouri: Missouri Geological Survey, Report of Investigation 46, 50 p.

Berry, A.W., 1970, Precambrian volcanic rocks associated with the Taum Sauk Caldera, St. Francois Mountains, Missouri (Ph.D. dissertation): University of Kansas, 147 p.

Berry, A.W., 1976, Proposed stratigraphic column for Precambrian volcanic rocks, western St. Francois Mountains, Missouri, *in* Studies in Precambrian Geology of Missouri: Missouri Geological Survey and Water Resources, Report of Investigations 61, p. 81-90.

Bickford, M.E., Van Schumus, W.R., and Zietz, I., 1986, Proterozoic history of the mid-continent region of North America: Geology, v. 14, p. 492-496.

Bickford, M.E., Lowell, G.R., Sides, J.R., and Nusbaum, R.L., 1977, Inferred structural configuration of the St. Francois Mountains Batholith, southeastern Missouri: Geological Society of America, Abstracts with Progams, v. 9, no. 5, p. 574-575.

Chapin, C.E., and Lowell, C.R., 1979, Primary and secondary flow structures in ash-flow tuffs of the Gribbles Run paleovalley, Central Colorado: Geological Society of America Special Paper 180, p. 137-154.

Chesner, Craig A., Rose, William I., and Hehanussa, Peter E., 1983, The Toba Caldera and Ignimbrite, Sumatra: Geological Society of America Abstracts with Programs, v. 15, no. 6, p. 543.

Cordell, L., 1979, Gravity and aeromagnetic anomalies over basement structure in the Rolla quadrangle and the southeast Missouri lead district: Economic Geology, v. 74, p. 1383-1394.

Dodge, F.C., 1979, The Uyaiiah ring structure, Kingdom of Saudi Arabia: U.S. Geological Survey Professional Paper, 774-E, 17 p.

Elston, W.E., 1984, Mid-Tertiary ash-flow tuff cauldrons, southwestern New Mexico: Journal of Geophysical Research, v. 89, B10, p. 8733-8750.

Hildebrand, R.S., 1984, Folded cauldrons of the early Proterozoic Labin Group northwestern Canadian Shield: Journal of Geophysical Research, v. 89, no. B10, p. 8429-8440.

Kisvarsanyi, E.B., 1980, Granitic ring complexes and Precambrian hot-spot activity in the St. Francois terrane, Mid-continent region, United States: Geology, v. 8, p. 43-47.

Kisvarsanyi, E.B., 1981, Geology of the Precambrian St. Francois Terrane, southeastern Missouri (Contribution to Precambrian Geology No. 8): Missouri Dept. of Natural Res., Div. of Geology and Land Survey, Report of Investigations 64, 60 p.

Lipman, P.W., 1976, Caldera-collapse breccias in the western San Juan Mountains, Colorado: Geological Society of America Bulletin, v. 87, p. 1397-1410.

Nusbaum, R.L., 1980, A Precambrian collapse caldera boundary in the St. Francois Mountains, southeast Missouri, (M.S. Thesis): University of Kansas, Lawrence, 77 p.

Nusbaum, R.L., 1981, A Precambrian collapse caldera boundary in the St. Francois Mountains, southeast Missouri: Geological Society of America, Abstracts with Programs, v. 13, p. 521.

Nusbaum, R.L., 1984, The Wah Wah Springs Tuff: Compositional variations in a dominant volume tuff (Ph.D. dissertation): University of Missouri-Rolla, 161 p.

Pratt, W.P., Anderson, R.E., Berry, Jr., A.W., Bickford, M.E., Kisvarsanyi, E.B., and Sides, J.R., 1979, Geologic map of exposed Precambrian rocks, Rolla $1^{0} \times 2^{0}$ quadrangle, Missouri: U.S. Geological Survey Miscellaneous Investigation Series, 1-1161.

Ratte, J.C., and Steven, T.A., 1967, Ash-flows and related volcanic rocks associated with the Creede Caldera, San Juan Mountains, Colorado: U.S. Geological Survey Professional Paper, 366, 81 p.

Sides, J.R., 1978, A study of the emplacement of a shallow granite batholith: The St. Francois Mountains, Missouri (Ph.D. dissertation): University of Kansas, Lawrence, 124 p.

Sides, J.R., Bickford, M.E., Shuster, R.D., and Nusbaum, R.L., 1981, Calderas in the Precambrian terrane of the St. Francois Mountains, southeastern Missouri: Journal of Geophysical Research, v. 86, n. B11, p. 10439-10364.

Silver, L.T., Conway, C.M., and Ludwig, K.R., 1986, Implications of a precise chronology for early Proterozoic crustal evolution and caldera formation in the Tonto Basin-Mazatazal Mounains region, Arizona: Geological Society of America Abstracts with Programs, v. 18, no. 5, p. 413.

Smith, R.L., 1979, Ash-flow magmatism: Geological Society of America Special Paper 180, p. 5-27.

Smith, R.L., and Bailey, R.A., 1968, Resurgent cauldrons, *in* studies in volcanology (Williams volume): Geological Society of America Memoir 116, p. 613-662.

Smith, R.L., Bailey, R.A., and Ross, C.S., 1961, Structural evolution of the Valles Caldera, New Mexico, and its bearing on the emplacement of ring dikes: U.S. Geological Survey Professional Paper 424D, p. D145-D149.

Thomas, J.J., Schuster, R.D., and Bickford, M.E., 1984, A terrane of 1350 to 1440 m.y.-old silic volcanic and plutonic rocks in the buried Proterozoic of the mid-continent and in the Wet Mountains, Colorado: Geological Society of America Bulletin, v. 95, p. 1150-1157.

Thompson, Margaret D., 1985, Evidence for a Late Precambrian caldera in Boston, Massachusetts: Geology, v. 13, p. 641-643.

GEOPHYSICAL STUDY OF A SMALL ULTRAMAFIC BODY NEAR NEWFOUND GAP BUNCOMBE COUNTY, NORTH CAROLINA[1]

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ABSTRACT

Gravity, ground magnetic, and radiometric surveying of a small ultramafic body near Newfound Gap in western North Carolina was undertaken to determine its three-dimensional extent and to evaluate the usefulness of these surveys in the study of ultramafic bodies. Gravity surveying yields positive anomalies when there is a high density contrast between the ultramafic material and the country rock. Density contrast is greatest for unaltered dunite because its common alteration products, serpentine and talc, have characteristically low density. Magnetite, a by-product of serpentinization of olivine, effectively increases the magnetic susceptibility of the dunite and increases the magnitude of magnetic anomalies associated with it. Therefore, these two types of geophysical surveys are complementary to one another when used to delineate dunite bodies such as this one.

Radiometric surveying, which measures gamma radiation, readily detects dunite where it has been emplaced in relatively highly radioactive country rock. However, this type of survey is effective only to a limited depth. Thus, blocks of ultramafic float which have migrated downslope will be detected by gamma ray surveys. As such, this method for delineating ultramafic bodies must be used with caution in areas of significant relief.

Gravity modeling of the anomaly over the small Newfound Gap ultramafite yields a body with maximum plan dimensions of 375 feet (115 meters) by 235 feet (72 meters) with a depth to the base of 330 feet (100 meters). Magnetic modeling indicates a body whose maximum dimensions are 345 feet (105 meters) by 230 feet (75 meters) with a depth to its base of 350 feet (107 meters). The small disparity in the dimensions derived from the two techniques is likely due to compositional variations within the dunite which alter the physical properties on which the models depend.

INTRODUCTION

In recent years, considerable renewed attention has focused on ultramafic bodies in the southern Appalachian Mountains because of their relevance to plate tectonic models and because of the commercial value of their highly refractory olivine. Ultramafic rocks in this region were studied collectively in a reconnaissance survey by Hunter (1941) and by Hunter and others (1942) for evaluation of their commercial olivine and chromite, respectively. Conventional field mapping as well as petrologic and geochemical studies were completed by several workers on selected ultramafic bodies in North Carolina (Kingsbury and Heimlich, 1978; Astwood and others, 1972; Carpenter and Phyfer, 1976; Hahn and Heimlich, 1977; Honeycutt and Heimlich, 1980; Dribus and others, 1982).

Until recently, however, little geophysical work has been done on them, and that which has been completed has yielded a much different distribution of ultramafic rock than was indicated previously on the basis of conventional field mapping alone (Dribus and others, 1982; Perez, 1979; Honeycutt and others, 1981; Schiering and others, 1982). A fairly extensive geophysical investigation of the Twin Sisters dunite in northern Washington was completed by Thompson and Robinson (1975). Their combined gravity and magnetic surveys

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enabled them to define the three-dimensional shape of that body and to detect a serpentine sheath surrounding it.

In the southern Appalachian region, where outcrops are sparse due to intense weathering, thick soil development, and extensive vegetative cover, additional tools are particularly necessary for adequate mapping of the many ultramafic bodies. Moreover, gravity and magnetic surveying produce information regarding three-dimensional size and shape, information which is practical from a commercial point of view. Radiometric surveying may also be useful in the surface mapping of poorly exposed ultramafic bodies.

This study was designed to determine the three-dimensional size and shape of a small ultramafic mass in western North Carolina using a variety of geophysical surveys and to evaluate the usefulness of these surveys in the study of ultramafic rock bodies generally. Detailed petrographic and x-ray analysis was done on samples of the ultramafite and the country rock to characterize variation in the body and to aid in evaluating densities and the magnetic characteristics of the rocks.

The ultramafic body investigated occurs approximately 14 miles (22 km) west of Asheville, within the Blue Ridge Province (Figure 1). The rocks crop out in the Newfound Gap area west of Newfound Road in southern Buncombe County, near the Buncombe County-Haywood County line, approximately four miles (6.4 km) northeast of Canton, North Carolina (Figure 2).

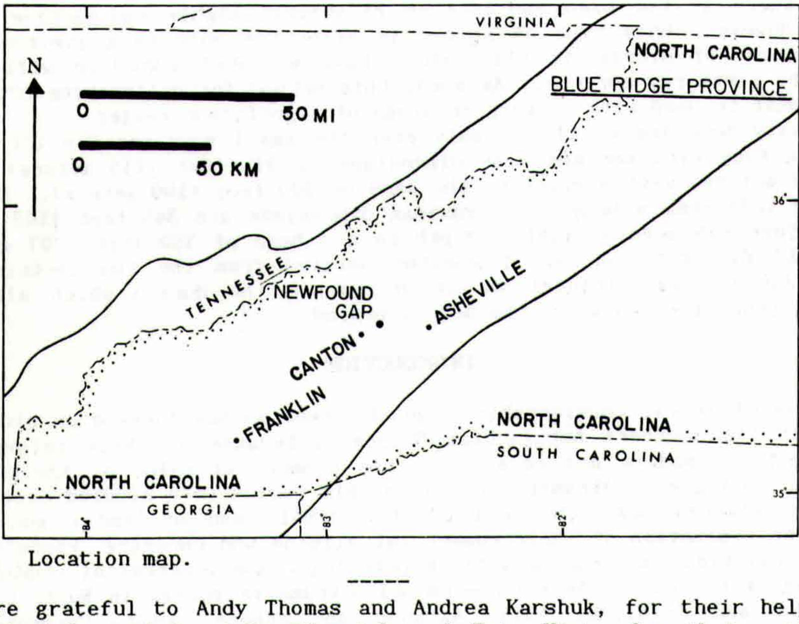

Figure 1. Location map.

We are grateful to Andy Thomas and Andrea Karshuk, for their help with the field work, and to John Plevniak and Eric Hirt, for their aid with computer programming. The cooperation of landowners Bill Boyd, Tom Silver, and Jack Brettler is greatly appreciated as well.

GEOPHYSICAL METHODS

Gravity and magnetic surveys were conducted following standard methods described in Dobrin (1976) and Telford and others (1976). Three measurements were made at gravity and magnetic stations using a Worden Gravity Meter and a Geometrics Model G-816 Portable Proton Precession Magnetometer, respectively. Precise elevations of gravity stations required for elevation and topographic corrections of gravity values were determined to within 0.1 foot, using a Wild T-2 Theodolite, and these surveys were tied to U.S. Geological Survey topographic quadrangle maps. Magnetometer stations were located by means of

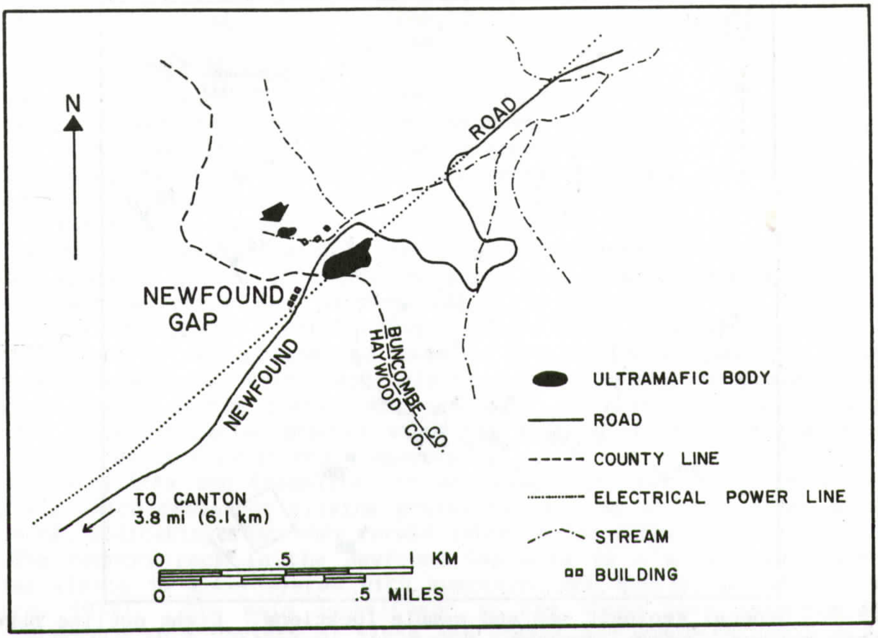

Figure 2. Newfound Gap location map; arrow identifies the ultramafic body studied.

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pace-and-compass traversing. Radiometric measurements were made with a Geometrics Model DISA-300 Portable Gamma Ray Spectrometer and were also located by pace-and-compass, following the approach used by Callahan and others (1978) and Schiering and others (1982).

The three types of surveys required establishing a station grid over the ultramafic mass and beyond it to obtain background measurements. However, locally, extensive vegetation or topographic obstacles prevented surveying small portions of the area. The gravity survey net was most limited because of the requirements for accurate elevation control by theodolite mapping.

Computerized linear regression was used to correct for drift effect relative to both gravity and magnetic values. Additional gravity corrections for elevation, latitude, large-scale topography (such as nearby mountains), the curvature of the earth, and the slab effect were completed using a U.S. Geological Survey program designed by Plouff (1977). The resulting Bouguer gravity data were plotted, contoured, and then modeled to determine the three-dimensional shape. Modeling programs used were based on equations by Hjelt (1972), for three-dimensional prismatic magnetic modeling, and those of Pant and Govindarajan (1979) and Telford and others (1976), for three-dimensional disc and two-dimensional prismatic gravity modeling, respectively.

Laboratory procedures included density measurements of dunite and country rock samples (using a Jolly Balance), measurement of the natural remanent magnetization direction (using a Fluxgate Magnetometer), and magnetic susceptibility measurements of dunite and country rock samples (using a Soiltest MS-3A Magnetic Susceptibility Bridge). Where thin-section analysis was conducted, modes were determined using a minimum of 1200 points and a grid designed to cover the entire thin section; analytical error was ±2 percent(Chayes, 1956). Plagioclase composition in the country rocks was determined by measuring the extinction angles of albite twins cut perpendicular to (010) and comparing them to the Michel-Levy determinative chart (Kerr, 1977). X-ray diffraction was employed for identification of the opaque minerals.

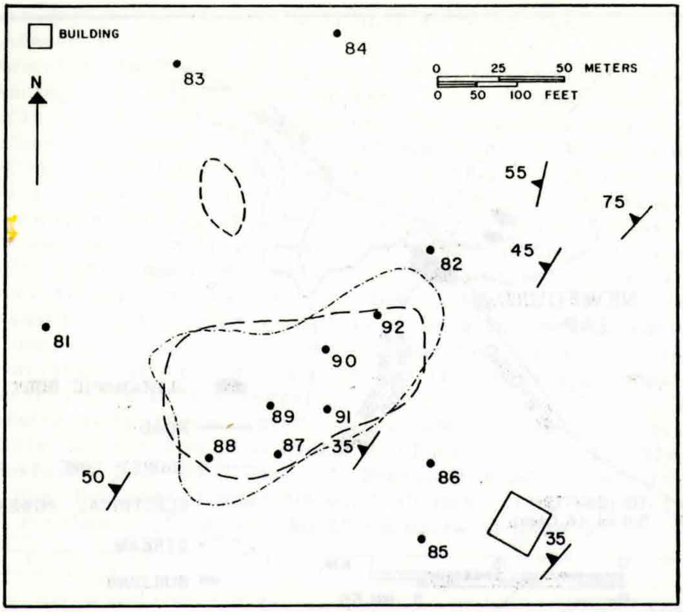

Figure 3. General geologic map and sample locations. Light outline based on gravity survey; heavy outline based on magnetic survey. A small serpentinized dunite mass, unconnected to the larger mapped body, occurs north of the main body.

PETROGRAPHY

Petrographic examination of dunite samples (Figure 3) reveals that the primary minerals are olivine and chromite (Table 1).

Table 1. Modal analyses of the dunite and country rock.

DUNITE SAMPLE NO.	**87**	**88**	**89**	**90**	**91**	**92**
Olivine	82.6	69.9	72.2	92.8	87.6	93.1
Serpentine	12.1	16.6	1.2	6.4	9.7	4.3
Chromite	2.3	1.2	1.1	0.6	1.5	1.6
Anthophyllite	0.2	1.1	2.3		0.1	0.8
Talc	2.3	11.2	23.0	0.2	1.1	0.1
Vermiculite	0.5					
Tremolite			0.2			
Chlorite	tr	tr	tr			

GNEISS SAMPLE NO.	**81**	**82**	**84**
Quartz	47.0	69.2	50.8
Plagioclase	10.6	12.0	25.2
K-feldspar		0.1	0.1
Muscovite	5.1	7.9	9.5
Biotite	35.3	7.6	12.0
Magnetite	1.3	1.2	0.7
Zircon, apatite, epidote	0.7	2.0	1.7
Plagioclase An	37	38	36

Olivine comprises 70-93% of the dunite with a mean value of 83%. Its Fo content averages 92%. In all samples, the olivine occurs typically as a mosaic of unstrained polygonal grains with numerous 120° triple-point junctions. Larger, irregularly shaped olivine porphyroclasts are distributed randomly throughout the rock; characteristically these grains show undulose extinction and strain-bands. Some olivine grains are highly fractured. Hydrous alteration is concentrated, to one degree or another, along these cracks or at grain margins.

Chromite, identified by x-ray diffraction, occurs as subhedral to anhedral disseminated grains which comprise from 0.6 to 2.3% of the rock (Table 1). Local alteration to chlorite occurs along grain boundaries.

Hydrous alteration minerals include serpentine (and associated secondary magnetite), anthophyllite, talc, chlorite, and locally, tremolite and vermiculite (Table 1). Serpentine occurs as fracture fillings in the olivine grains and as irregular patches replacing the borders of some olivine grains. Minute anhedral grains of secondary magnetite occur disseminated throughout the serpentine veins and patches, as a by-product of the alteration of olivine to serpentine. Talc and vermiculite occur commonly as flaky masses which replace olivine grains and they are locally intergrown with serpentine. In several instances, talc completely surrounds the olivine grains, such that only the center of the grain remains intact. Chlorite is present in several samples as an alteration product along the margins of chromite grains. It is typically present only in trace amounts.

Anthophyllite and tremolite (in one sample) occur typically as acicular crystals which cross-cut olivine grains as well as all the other alteration products, indicating that they formed later in the alteration history.

The country rock in the Newfound Gap area is a mica-plagioclase-quartz gneiss (Table 1) interlayered with muscovite and biotite schist (Palmer and others, 1977). The main constituent in the three thin sections studied, quartz occurs in a variety of sizes and shapes and exhibits undulose extinction. It is typically xenoblastic and irregularly fractured throughout.

Generally slightly larger than the quartz grains, plagioclase exhibits albite twinning; its average composition is An37. Microcline, displaying grid-type polysynthetic twinning, is present in minute amounts.

Muscovite and biotite, locally with quartz inclusions, occur as bent lath-shaped crystals which form semi-parallel layers within the gneiss. Both occur also as flaky aggregates. Non-opaque accessory minerals include very small amounts of zircon, apatite, and epidote. Zircon and apatite are associated with the mica crystals. Magnetite occurs mainly as minute anhedral grains, along cleavage traces in the micas, especially biotite, and as larger, ragged interstitial grains.

GRAVITY SURVEY

Figure 4 shows the location of surveyed gravity stations. The layout of the survey did not cross the boundaries of the body to the northeast or southwest so that modeling yields a minimum length for the body. Contoured Bouguer gravity values (Figure 5) yield a gravity anomaly which indicates that the long axis of the body is oriented ENE at an angle of approximately 25 degrees to the well defined northeast-southwest strike of foliation in the country rock (Figure 3).

Rather than closing over the body, the gravity contours have an oblong shape that pinches in the middle, suggesting the presence of less material in the center of the body than at the ends (Figure 5). The high gravity values located toward the ends of the body are interpreted as representing a combination of more mass (thicker and larger in areal extent) and possibly a greater density contrast between the dunite and surrounding gneiss than at the center.

In association with the gravity survey, density measurements were made for six dunite and six gneiss samples (Figure 3 and Table 2). Mean density for the dunite (3.23 g/cm3) is significantly higher than that for the gneiss (2.60 g/cm3).

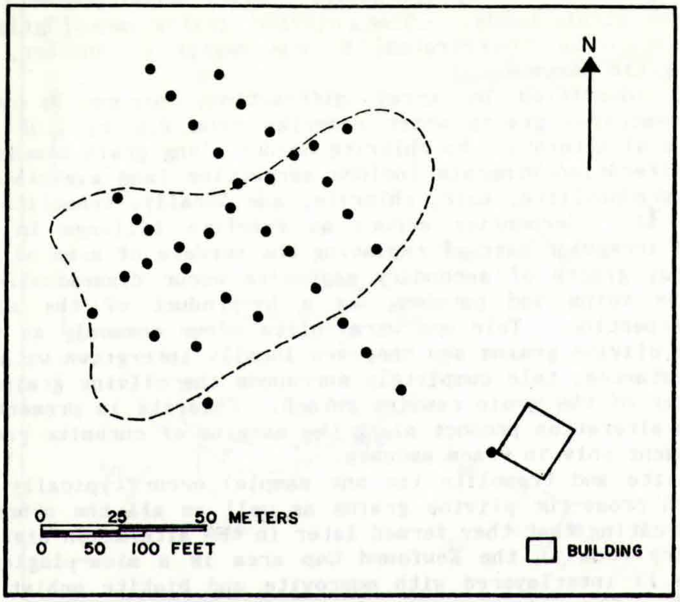

Figure 4. Location map for gravity survey stations. Survey lines were restricted by dense vegetation which prevented accurate measurement of elevations over the northeast and southwest margins of the body.

The data were modeled using a program based on equations by Pant and Govindarajan (1979) which employs a series of stacked vertical discs in which the radius, thickness and depth of the discs can be varied. The resulting three-dimensional model consists of six different discs of various sizes, thicknesses, and density contrasts (Figure 6), with the dashed discs at depth and solid discs at the surface.

Table 2. Density and magnetic susceptibility measurements for the dunite.

SAMPLE NO.	DENSITY (g/cm3)	SUSCEPTIBILITY (x 10-5 cgs)	MODAL PERCENT SERPENTINE
Dunite			
87	3.14	11.15	12.1
88	3.23	31.54	16.6
89	3.30	5.51	1.2
90	3.25	9.85	6.4
91	3.24	8.75	9.7
92	3.22	7.85	4.3
Mean	3.23	12.44	8.4
Standard Deviation	0.05	9.55	5.6
			MODAL PERCENT MAGNETITE
Gneiss			
81	2.65	2.68	1.3
82	2.67	2.03	1.2
83	2.62	1.73	
84	2.57	1.34	0.7
85	2.46	0.75	
86	2.60	0.67	
Mean	2.60	1.75	1.1
Standard Deviation	0.08	0.84	0.3

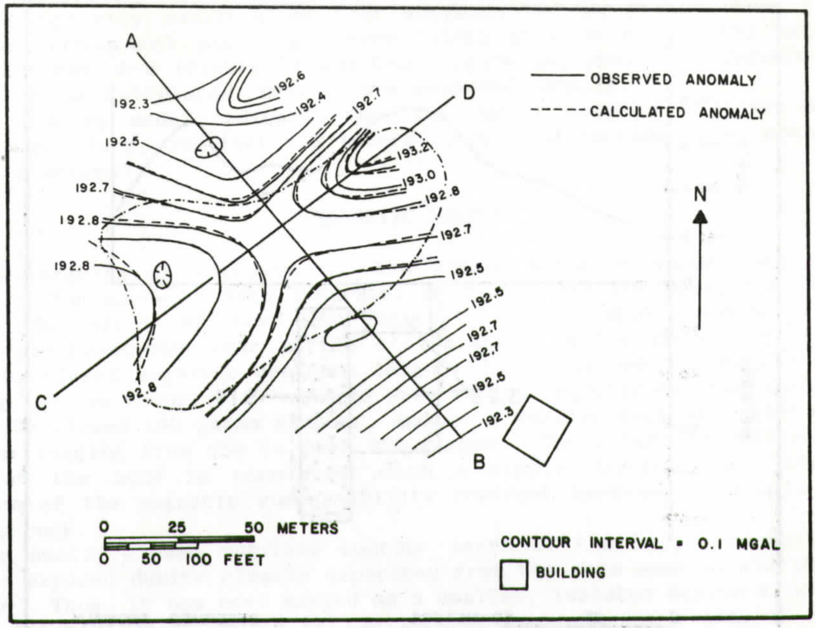

Figure 5. Contoured gravity map.

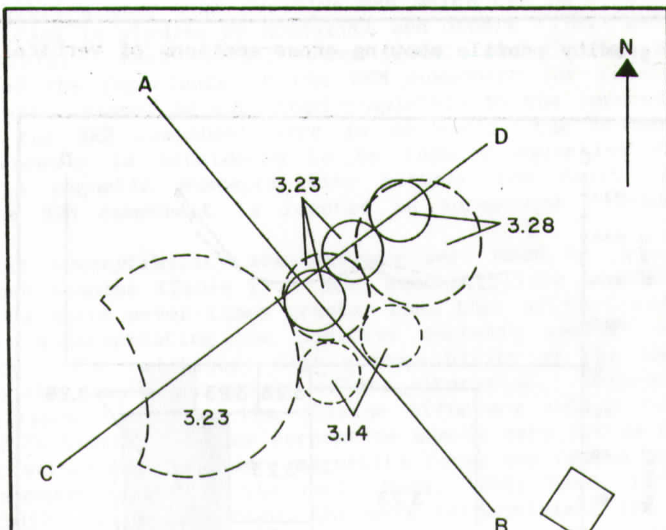

Figure 6. Map showing the configuration of stacked discs for the gravity model. Dashed discs are lower in the stacking than the solid discs. The largest disc on the southwest end of the body was modified by a series of cylindrical cut-outs to improve the fit between the calculated and observed data. Densities, indicated for each disc, contrast with the background density of 2.64 g/cc for the gneiss closest to the dunite.

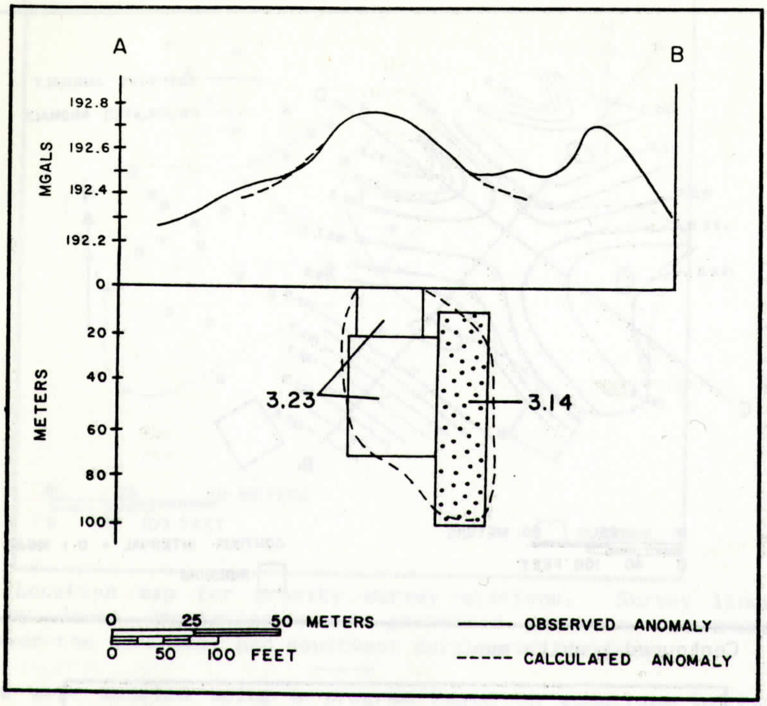

Figure 7. A-B gravity profile showing cross-sections of vertical discs with densities.

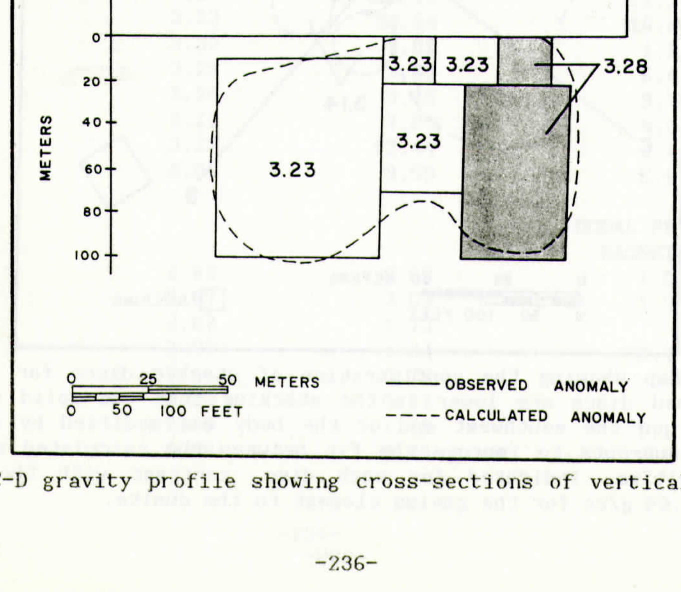

Figure 8. C-D gravity profile showing cross-sections of vertical discs with densities.

The largest disc at the southwest end of the model was modified by a cut-out to improve the match between the observed and calculated curves. It is the need for this modification that suggests to the authors that the body does not extend much past the survey points at this end of the body. Two cross-sections, A-B (Figure 7) and C-D (Figure 8), show the subsurface shape and fit of the calculated curve to the observed anomaly.

The gravity model indicates that the dunite body is at least 375 feet (115 meters) long, 235 feet (72 meters) wide, and extends to a depth of 330 feet (100 meters).

MAGNETIC SURVEY

The location of stations for the ground magnetic survey are given in Figure 9. The magnetic data indicate, as do the gravity data, that the long axis of the body is oriented ENE (Figure 10) at an angle of about 30 degrees to the northeast-southwest strike of foliation in the country rock (Figure 3). The closed magnetic contours (Figure 10) encompass all dunite outcrops and suggest two things: 1) a large area of low magnetic contrast, represented by the closed 100 gamma contour, and 2) a smaller area of higher magnetic contrast ranging from 150 to over 300 gammas. The closed depression contour north of the body is consistent with a simple dipolar field that is a function of the magnetic susceptibility contrast between the dunite and the country rock.

The small, closed positive contour north of the body encloses a small area of exposed dunite clearly separated from the main mass by the depression contour. Thus, it has been mapped as a smaller, isolated dunite mass.

Qualitative measurements of the natural remanent magnetization direction were made for two oriented samples (numbers 88 and 91, Figure 3). The samples exhibit a very weak positive NRM which can be considered negligible, as was the case in studies by Honeycutt and others (1981) and by Thompson and Robinson (1975). Beck (1975), Green (1960), and Girdler and Peter (1960) demonstrated the importance of the NRM component for interpreting magnetic anomalies which cannot be explained completely by the induced magnetic field, but since the NRM component here is so small, the primary cause of the magnetic anomaly is considered to be induced magnetism reflected by the contrast in magnetic susceptibility between the dunite and the country rock. The NRM component is ignored in subsequent considerations of the data.

Magnetic susceptibility measurements were made for six dunite and six country rock samples (Table 2). Mean susceptibility of the dunite (12.44 x 10-5 cgs) is about seven times greater than that of the country rock (1.75 x 10-5 cgs), substantiating the positive magnetic anomaly located over the dunite body. The relatively high susceptibility of the dunite is directly related to the nature and extent of its alteration. Serpentinization of the dunite releases iron from the olivine structure (which typically contains 8-10 mole% Fe_2SiO_4). Because serpentine admits very little of this iron into its atomic structure, secondary magnetite forms and causes an increase in the magnetic susceptibility of the rock (Hess, 1933; Saad, 1979; Honeycutt and others, 1981). Generally then, the more serpentinized the ultramafic rock, the higher the susceptibility values should be, and this correlation is seen in Table 2 between modal percent serpentine and the mean value of susceptibility.

Gneiss susceptibility values are significantly lower than those of the dunite due to the correspondingly lower modal percent magnetite in the gneiss compared to the serpentinized dunite (Table 2). The cause of the magnetic anomaly, then, is considered to be the contrast in susceptibility values reflected by the contrast in modal percent magnetite in the dunite and gneiss.

The profile A-B (Figure 10) was chosen for modeling by three-dimensional prisms because it perpendicularly bisects the area of dense contours over the

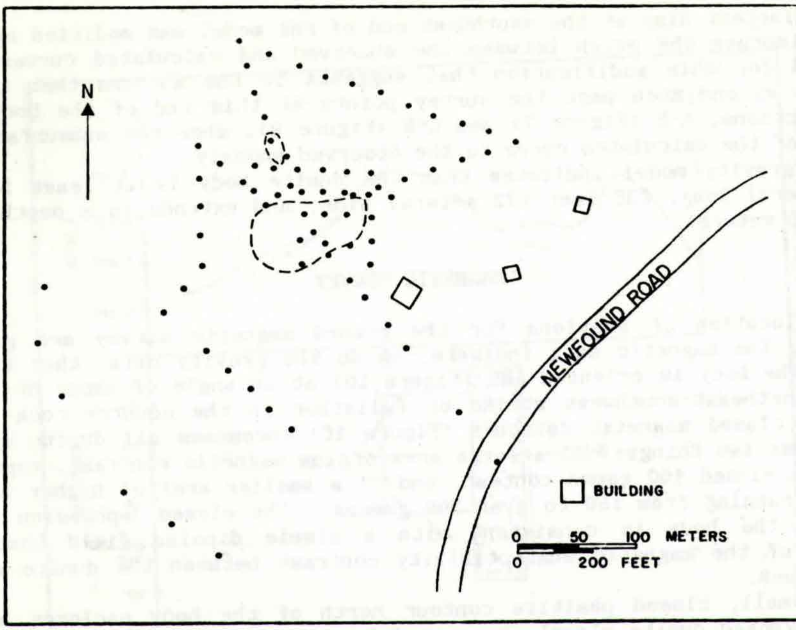

Figure 9. Location map for magnetic stations.

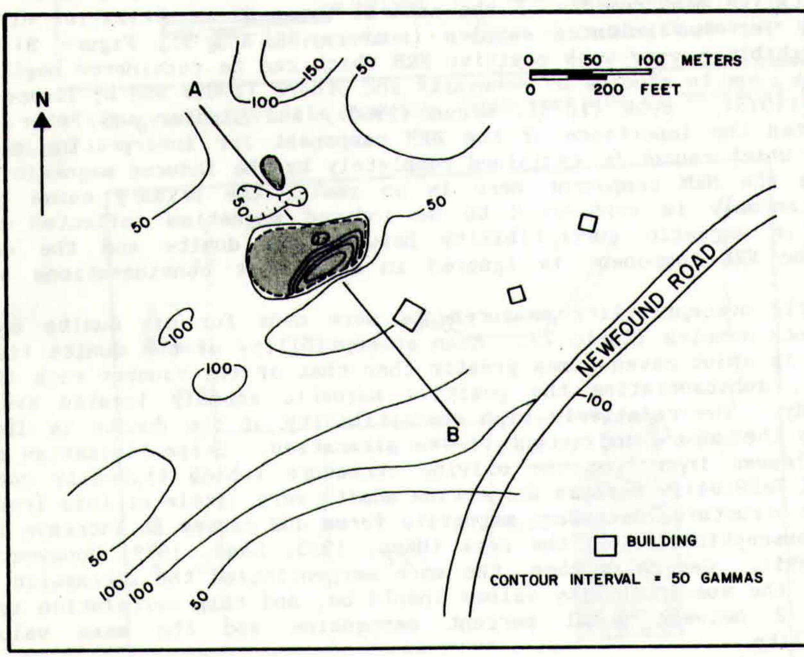

Figure 10. Contoured magnetic map.

ultramafic body. The observed anomaly consists of two components: a narrow, high peak (concentrated area of high susceptibility) overlapping a broad, relatively lower curve (larger area of relatively lower susceptibility) (Figure 11). The two portions were individually modeled, using a program based on the equations of Hjelt (1972), with the resulting curve representing the addition of the two curves. The model is dominated by two nearly vertical prisms which exhibit relatively higher and lower susceptibilities. A third, shallow horizontal prism was introduced so that the model would remain consistent with the outcrop data as well as the magnetic survey.

Magnetic data showed a small high to the north of the main body (Figure 10) which was not expressed in the gravity data. This high is coincident with a small outcrop of dunite (Figure 3) and indicates that the underlying body is small, shallow, rather serpentinized, and is completely unattached to the larger modeled mass.

Thus, the magnetic survey indicates that the body has maximum plan dimensions of 345 feet (105 meters) by 230 feet (70 meters), and that it extends to a maximum depth of 350 feet (107 meters). The dimensions are very similar to those obtained from the gravity model (115 meters by 72 meters by 100 meters deep). The general plan shapes of the two models are very similar although the orientation of its long axis is slightly different and there is a 10% difference between the lengths of the two (Figure 12). Figure 13 provides a comparison of the gravity and magnetic profiles for cross-section A-B.

DISCUSSION OF GRAVITY AND MAGNETIC SURVEY RESULTS

Differences in results between gravity and magnetic surveys are a function of the physical property (density versus magnetic susceptibility) which each of these measures. For dunite bodies in North Carolina, density contrasts are highest where the rock is unaltered. Here, the high density of the olivine results in an overall high density of the dunite. High magnetic contrasts, however, require anywhere from moderate to extensive alteration of the olivine to serpentine, because the magnetite which forms as a by-product of this alteration must be present in order to yield susceptibility contrasts of large enough magnitude to be detected. Serpentine, however, has a very low density (2.55 - 2.60 g/cm3) (Deer and others 1966), which results in a lowering of the amplitude of a gravity anomaly.

The two types of geophysical surveys complement one another; that is, the less altered the material is, the higher the density (gravity anomaly) and the lower the susceptibility (magnetic anomaly) will be. Conversely, an extensively altered dunite body will possess a low overall density and a very high average susceptibility. This is true not only between dunite masses, but also within a single mass. As shown in Figure 5, a high density contrast exists at the ends of the dunite studied and is reflected by gravity highs there. However, the susceptibility contrast is relatively low at the same locations (Figure 10). Similarly, in the southeastern portion of the body, where high magnetic values indicate higher susceptibility contrast, a corresponding low density contrast and weak gravity anomaly exist.

Geologically, we interpret this to reflect greater alteration of the body (with correspondingly low gravity and high magnetic susceptibility values) along its southeastern portion but with alteration almost absent at its ends. Petrographic analyses of the dunite samples support this observation. Sample 92, located at the northeast end of the body (Figure 3), shows little alteration (4.3 modal percent serpentine) (Table 1), but density contrast is very high in this area. Samples 87, 88, and 91 (Figure 3), located in the south and southeastern portion (where density contrast is low and magnetic susceptibility contrast is high), are more extensively altered (12.1, 16.6, and 9.7 modal percent serpentine, respectively) (Table 1). The disparity in depths to the bottom of the dunite, based on the gravity and magnetic models, suggests that a thin sheath of highly altered dunite (seven meters thick) may extend beneath the body in the southeast.

Clearly, the geophysical methods detect different aspects of the rock, due to the different physical properties measured by each. As shown here, a combination of the two proves very useful in three-dimensional modeling of this ultramafic mass.

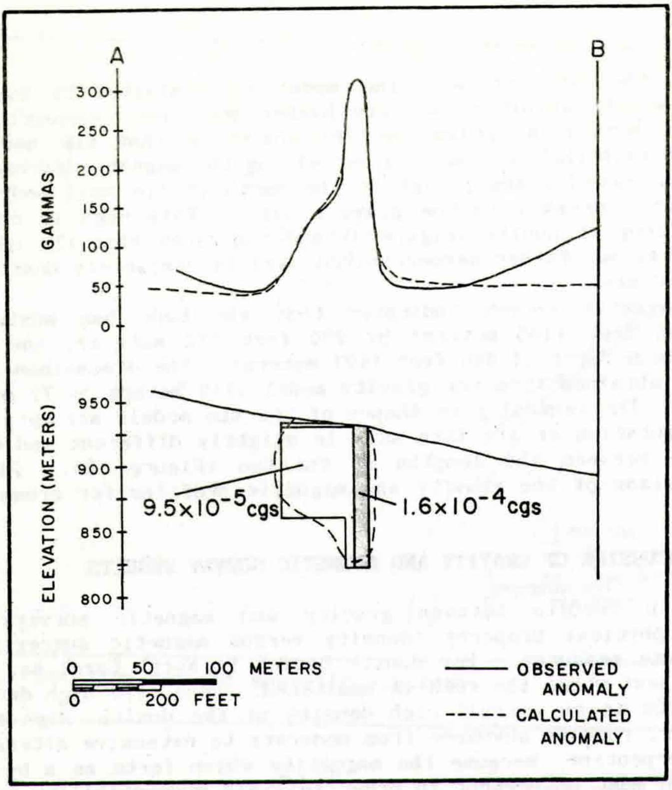

Figure 11. A-B magnetic profile showing cross-section of 3-dimensional vertical prisms with susceptibilities. Background susceptibility for the gneiss is 1.75 x 10-5 cgs.

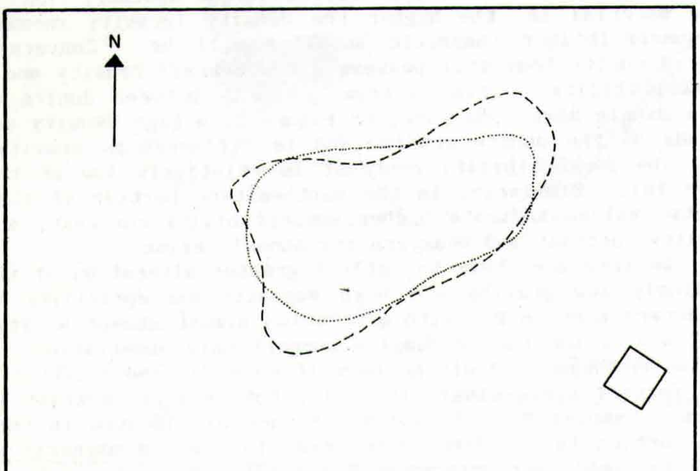

Figure 12. Comparative gravity (dashed line) and magnetic (dotted line) survey outlines of the dunite body.

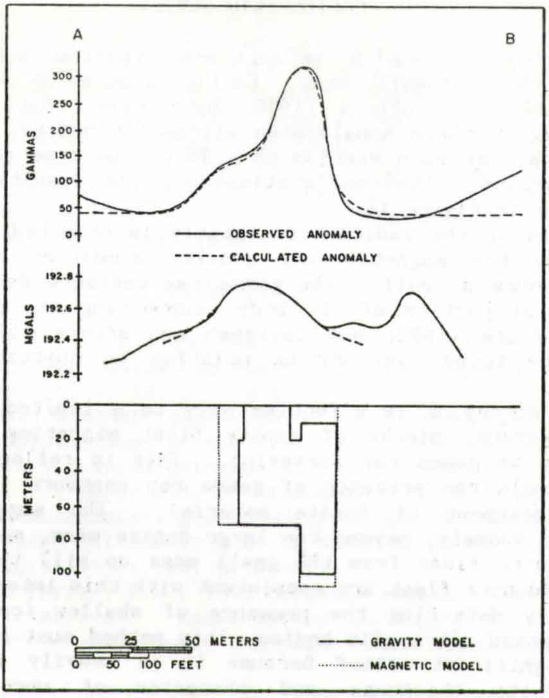

Figure 13. Comparison of gravity and magnetic profiles along line A-B for the dunite. That part of the body on the right has lower density and higher susceptibility indicating that it has been more serpentinized than the volume of higher density and lower susceptibility on the left side of the figure.

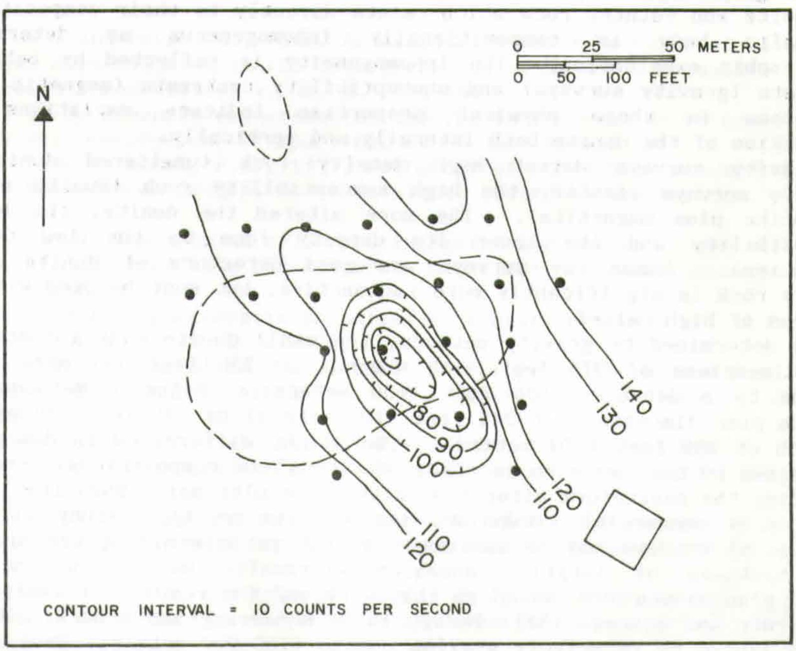

Figure 14. Location of gamma ray stations and contoured radiometric map.

RADIOMETRIC SURVEY

Gamma radiation, produced by potassium-40, thorium, and uranium, was also used to outline the ultramafic body. Earlier studies by Schiering and others (1982) and Callahan and others (1978) have shown that the method may be useful in mapping southern Appalachian ultramafic bodies. Three radiometric readings were taken at each station on a 20-second time constant and reduced to counts per second. Station locations and the resulting contoured gamma ray map are shown in Figure 14.

The long axis of the radiometric anomaly is oriented at right angles to that defined by the magnetic and gravity anomalies and normal to the topographic contours as well. The gamma ray contours decrease in magnitude toward the central portion of the body, supporting the previous results of Schiering and others (1982) and Callahan and others (1978) that generally lower values are found over dunite relative to quartzofeldspathic gneiss country rock.

Radiometric surveying is effective only to a limited depth of 0.3 to 1 meter, and therefore, blocks of dunite float migrating downslope will be detected readily by gamma ray surveying. This is reflected in the present study which reveals the presence of gamma ray contours following faithfully the downslope movement of dunite material. The suggested northwestern extension of the anomaly, beyond the large dunite mass, may reflect downslope migration of dunite float from the small mass up hill to the north. Field observations of dunite float are consistent with this interpretation.

While readily detecting the presence of shallow (concealed) dunite in southern Appalachian ultramafic bodies, this method must be used with caution in areas of significant relief because it is heavily influenced by slope movements and the thickness and character of unconsolidated surface materials.

CONCLUSIONS

The geophysical surveys used in this study detect physical properties in the dunite and country rock which relate directly to their composition. The ultramafic body is compositionally inhomogeneous as determined by petrographic examination. The inhomogeneity is reflected by both density contrasts (gravity surveys) and susceptibility contrasts (magnetic surveys). Variations in these physical properties indicate variations in the composition of the dunite both laterally and vertically.

Gravity surveys detect high density rock (unaltered dunite) while magnetic surveys identify the high susceptibility rock (dunite altered to serpentine plus magnetite). The more altered the dunite, the higher its susceptibility and the lower its density (due to the low density of serpentine). Gamma ray surveys are good detectors of dunite where the country rock is significantly more radioactive, but must be used with caution in areas of high relief.

As determined by gravity methods, the small dunite body investigated has plan dimensions of 375 feet (115 meters) by 235 feet (72 meters) and it extends to a depth of 330 feet (100 meters). Magnetic methods indicate maximum plan dimensions of 345 feet (105 meters) by 230 feet (70 meters) and a depth of 350 feet (107 meters). The slight differences in dimensions, as determined by the two methods, is a result of the compositional inhomogeneity including the nonuniform alteration within the ultramafic mass itself.

From a commercial viewpoint, the results of this study suggest that geophysical surveys may be extremely useful in determining the areal extent and thickness of largely concealed ultramafic bodies in the southern Appalachian Mountains. Based on this work and the results of earlier studies (Honeycutt and others, 1981; Perez, 1979; Schiering and others, 1982), these bodies extend to relatively shallow depths (100-200 meters). Thus it may not be feasible to mine on a large scale many of the bodies which crop out in this region.

Magnetic surveys appear to be the most useful because they are less time-consuming than gravity surveys, are not as hindered by the presence of thick vegetation, require less elaborate data reduction, and appear to agree well with available outcrop information. However, magnetic surveys are effective only for serpentinized dunite bodies, some of which may be so altered that they are not worth mining. Unaltered dunite, of great commercial value, would go undetected by magnetic surveys, yet would be easily detected by gravity surveys. Here, unaltered dunite is most easily detected due to its relatively high density, but with increasing amounts of alteration of the dunite, gravity surveying becomes less useful.

REFERENCES CITED

Astwood, P.M., Carpenter, J.R., and Sharp, W.E., 1972, A petrographic study of the Dark Ridge and Balsam Gap dunites, Jackson County, North Carolina: Southeastern Geology, v. 14, p. 183-194.

Beck, M.E., Jr., 1975, Remanent magnetism of the Twin Sisters dunite intrusion and implications for the tectonics of the western Cordillera: Earth and Planetary Science Letters, v. 26, p. 263-268.

Callahan, J., Hearn, F., Heron III, S., Schumaker, D., and Swanson, S., 1978, Utilizing soil geochemistry, ground magnetic and radiometric surveys to outline a Blue Ridge ultramafic body in North Carolina: Journal of Geochemical Exploration, v. 10, p. 53-61.

Carpenter, J.R., and Phyfer, D.W., 1976, Olivine compositions from southern Appalachian ultramafics: Southeastern Geology, v. 16, p. 169-172.

Chayes, F., 1956, Petrographic modal analysis: John Wiley and Sons, New York, 113 p.

Deer, W.A., Howie, R.A., and Zussman, J., 1966, An introduction to the rock-forming minerals: Longman Group Limited, London, p. 242-249.

Dobrin, M.B., 1976, Introduction to Geophysical Prospecting: McGraw-Hill, New York, 630 p.

Dribus, J.R., Heimlich, R.A., and Palmer, D.F., 1982, The "Deposit No. 9" dunite, Macon County, North Carolina: Southeastern Geology, v. 23, p. 51-67.

Girdler, R.W., and Peter, G., 1960, An example of the importance of natural remanent magnetization in the interpretation of magnetic anomalies: Geophysical Prospecting, v. 8, p. 474-484.

Green, R., 1960, Remanent magnetization and the interpretation of magnetic anomalies: Geophysical Prospecting, v. 8, p. 98-110.

Hahn, K.R., and Heimlich, R.A., 1977, Petrology of the dunite exposed at the Mincey Mine, Macon County, North Carolina: Southeastern Geology, v. 19, p. 39-53.

Hess, H.H., 1933, The problem of serpentinization and the origin of certain chrysotile asbestos talc and soapstone deposits: Economic Geology, v. 28, p. 634-657.

Hjelt, S.E., 1972, Magnetostatic anomalies of dipping prisms: Geoexploration, v. 10, p. 239-254.

Honeycutt, F.M., and Heimlich, R.A., 1980, Petrology of the Balsam Gap dunite, Jackson County, North Carolina: Southeastern Geology, v. 21, p. 251-260.

_________, Heimlich, R.A., and Palmer, D.F., 1981, Magnetic survey of the Balsam Gap dunite, Jackson County, North Carolina: Southeastern Geology, v. 22, p. 223-231.

Hunter, C.E., 1941, Forsterite olivine deposits of North Carolina and Georgia: North Carolina Department Conservation and Development,Bulletin 41, 117 p.

________, Murdock, T.G., and McCarthy, G.R., 1942, Chromite deposits of North Carolina: North Carolina Department of Conservation and Development, Bulletin 42, 39 p.

Kerr, P.F., 1977, Optical mineralogy: McGraw-Hill, Inc., New York, 492 p.
Kingsbury, R.H., and Heimlich, R.A., 1978, Petrology of an ultramafic body near Micaville, Yancey County, North Carolina: Southeastern Geology, v. 20, p.33-46.
Palmer, D.F., Heimlich, R.A., and Kolb, R.J., 1977, Petrofabric analysis of the Newfound Gap dunite, Haywood and Buncombe Counties, North Carolina (Abs): Geological Society of America Abstracts with Programs, v. 9, p. 640.
Pant, M.M., and Govindarajan, J., 1979, Computation of gravity effects of two- and three-dimensional bodies with radial symmetry: Computers and Geosciences, v. 5, p. 313-323.
Perez, E.D., 1979, Magnetic surveys of selected ultramafic deposits in western North Carolina: Unpublished M.S. thesis, Kent State University, 43 p.
Plouff, D., 1977, Preliminary documentation for a FORTRAN program to compute gravity terrain corrections based on topography digitized on a geographic grid: United States Geological Survey Open File Report 77-535, 45 p.
Saad, A.H., 1969, Magnetic properties of ultramafic rocks from Red Mountain, California: Geophysics, v. 34, p. 974-987.
Schiering, M.H., Heimlich, R.A., and Palmer, D.F., 1982, Radiometric and magnetic surveys of the Dark Ridge dunite, Jackson County, North Carolina: Southeastern Geology, v. 23, p. 89-98.
Telford, W.M., Geldart, L.P., Sheriff, R.E., and Keys, D.A., 1976, Applied Geophysics: Cambridge University Press, New York, p. 72-74.
Thompson, G.A., and Robinson, R., 1975, Gravity and magnetic investigation of the Twin Sisters dunite, northern Washington: Geological Society of America Bulletin, v. 86, p. 1413-1422.